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Tectonic Regimes

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ABSTRACT

The concept of tectonic regimes is developed so as to specify the symmetry and orientation of bulk deformation for use in tectonic analysis. Four symmetry patterns combined with possible arrangements of principal axes in vertical and horizontal directions as in a compression belt give thirteen regimes which are in turn defined and tentatively related to some known tectonic structures. A theoretical orogenic sequence of regimes is developed and the paper concludes with a review of some of the difficulties.

I. INTRODUCTION

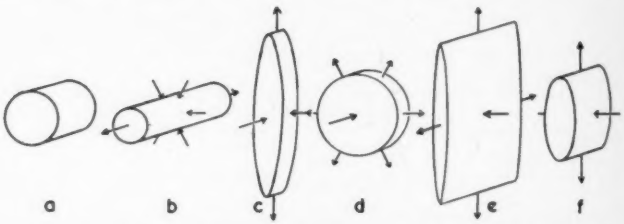
THIS paper is designed to illustrate a method of tectonic analysis briefly referred to under the heading " Tectonic Orientation " in an earlier paper (Harland, 1956). The concept of a tectonic regime is intended to define the nature and orientation of bulk strain during tectonic deformation.

Because of rock inhomogeneity, a single simple large-scale movement may generate a variety of minor structures (see Text-fig. 3). Each minor structure may be accommodated in the same overall movement. It is to designate the movement pattern common to all the structures that the term " tectonic regime " is used. Examples of the kind of pattern envisaged are shown in Text-fig. 1. In this figure, *b* and *c* show the same patterns but belong to different regimes, similarly *d* and *e*, for the concept is intended to imply an orientation as well as a pattern.

A total deformation as observed may be the result of a succession of movements in different directions and each transient movement pattern is distinguished as a separate regime. At any one time during an orogeny, then, different regimes may be in operation in different places, and in any one place a series of regimes may operate in succession. The concept of a tectonic regime makes discussion of tectonic events easier both by specifying a number of tectonic situations and by relating diverse structures to any one situation.

For each movement pattern a stress field may be guessed. Quantitative knowledge of this for past tectonic events may be unobtainable but at least the orientation and relative magnitude of the principal stresses may in some cases be inferred. Until the relationship in various circumstances between stress and strain is known it is proposed to specify the regime by the movement (or resulting strain) pattern. Ultimately some conceptual dichotomy of stress and strain regimes may be necessary but for the present a simple relationship is in mind and the strain regime carries with it the implication of the corresponding stress regime in ideal circumstances.

In the specification of a regime either by movement or by stress, the arrangement of relative magnitudes and the orientation of the arrangement are independent. It will be argued that in conjunction



TEXT-FIG. 1.—Representations of some movement patterns by the deformation of a cylinder (*a*) as follows:—
(*b*) and (*c*) Similar movement pattern differing only in orientation of elongation.
(*d*) and (*e*) Similar movement pattern differing only in orientation of flattening.
(*f*) Example of pattern of movement in one plane.

with four arrangements the introduction of vertical and horizontal co-ordinates generates thirteen separate regimes.

II. Symmetry of Bulk Deformation

Deformation of an ideal isotropic medium under extended stress is generally such that a sphere becomes an anisometric ellipsoid. However, there are possible theoretical cases where equalities of stress produce analogous equalities in strain, and the sphere becomes oblate, prolate, or simply a sphere of different size. The principal stresses in these cases are:—

(i) All equal.

(ii) One major and two equal minor.

(iii) Two equal major and one minor.

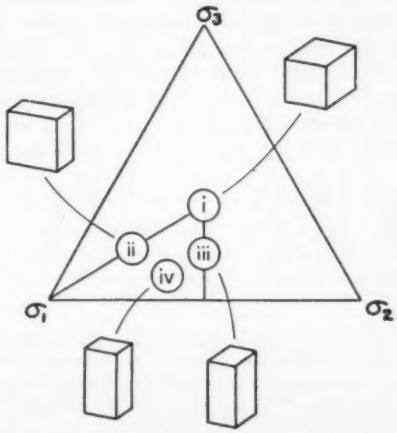
(iv) All unequal (the general case).

By comparing the strain ellipsoid with the indicatrix of optical mineralogy, these cases may be given names for reference; and the movements typical of the four systems are also summarized thus:—

TABLE 1

Stress pattern.	*Name.*	*Movement pattern.*
(i) All equal . .	isotropic	overall compression or dilatation.
(ii) 1 major, 2 minor .	uniaxial — ve	radial outflow (axial compression).
(iii) 2 major, 1 minor .	uniaxial + ve	axial extension (radial compression).
(iv) All unequal . .	biaxial	mainly two-dimensional flow.

Note.—The use of signs may be extended to the biaxial case by exact analogy if desired.



TEXT-FIG. 2.—Stress triangle with apices representing the three principal stresses and part of the triangle showing the position of the four symmetry types of Table 1. Each is illustrated by a cube deformed according to such a stress field.

These four bulk symmetry groups may be illustrated (as in Text-fig. 2) either as derivatives of an isometric solid or as fields in a triangular diagram in which the three co-ordinates are the principal stresses.

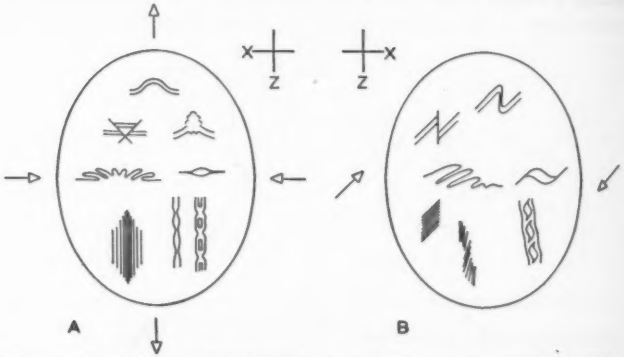
For the application of this classification to actual occurrences, modifications must be made on two counts:—

(1) Exact equality is theoretical: practical classification must be based on approximate equality only, whether of observed strain or of inferred stress.

(2) Actual rock bodies are not homogeneous and can sustain shear.

The first consideration gives reality to the three special cases (i), (ii), and (iii). If these are made to include examples of *approximate* equality of stresses, then the examples which remain of case (iv) are only those which have their intermediate stress well centred between the maximum and the minimum. This may be a necessary condition for flow to be restricted to two dimensions.

The second consideration introduces several complications. Firstly, the inhomogeneity provokes variety in the reaction to stress. Text-fig. 3 shows how different rock units may accommodate the same stress. A consequence is that in attempts to recognize a stress system from observations of actual structures, deformation must be con-



TEXT-FIG. 3.—Diagram showing variety of structures accommodated within the same bulk movement pattern. The derivation (A) shows symmetrical structures and (B) shows assymmetrical or predominantly shear structures. Both give the same bulk effect and all structures as illustrated belong to the same regime.

sidered on the same scale as the stress system. That this scale must be large enough to ensure conformity of the stress and strain patterns has been specified in introducing the concept. The total effective deformation must, therefore, be estimated by "bulking" all the smaller structures and noting their overall, not their individual, effects. In this process a mass of monoclinic details, for example, may often be reduced to an orthorhombic bulk change and some structural elements which give distinctive symmetry, such as foliation and lineation, may be excluded from the consideration of regimes unless the movements they reflect are significant. Thus, mechanical or other (e.g. apparent) inhomogeneity may give structures a lower (apparent) symmetry than the stresses producing them; in particular, even an orthorhombic stress field may give monoclinic or triclinic structures.

When transmission of shearing stress is admitted a variety of

predominantly shear structures may be generated; but *because* these may be accommodated in an ellipsoid derived from a sphere and having the same shape and orientation as the one representing the other mode of deformation, both are taken to represent the same regime. Such a contrast is made between Text-figs. 3*a* and *b*.

Attempts to recognize a bulk pattern from observations may thus suffer from a variety of difficulties. At depth where all stresses are big, deformation is plastic and the effect of rock inhomogeneity is minimized, " bulking " is simple and the net extensions and compressions may even be fully represented in the fabric of a simple hand-specimen. However, in a succession of movements only the net change or the last one may be detectable. On the other hand, nearer the surface, structures are more varied and " bulking " is more difficult; but overall deformation if it has taken place *is* detectable and the possibility of disentangling successive regimes remains.

For the immediate development of the regime concept, systems on a crustal scale are considered, where " bulking " is relied upon to reduce discrepancies of symmetry between observed deformation and an inferred stress pattern.

III. Orientation of Bulk Deformation

Hitherto we have considered the symmetry of bulk deformation. To complete the development of the regime concept we must now consider the orientation in the earth's crust of the distinctive strain (or stress) axes.

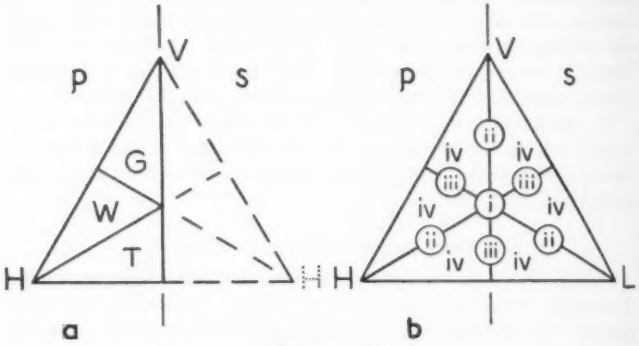
The vertical direction, explicitly or implicitly, permeates all tectonic description as, for instance, in the use of regional co-ordinates. For another instance we may recall Anderson's three conditions of faulting (1951). The erection of three classes of fault, thrust, wrench, and gravity [1] according as the vertical stress is minimum, intermediate, or maximum, assumes that the vertical is the approximate direction of one principal stress. The assumption is justified if a sufficiently large segment of the crust be considered though boundary conditions may modify this assumption (cf. Hafner, 1951).

These fault types, then, may be considered as particular expressions of three general situations which may have generated folding and cleavage as well as faulting. So an extension of the nomenclature to regimes may be considered; hence the proposed use of thrust, wrench, and gravity regimes.

The three cases may be conveniently shown on a triangular diagram

[1] " Gravity " is used here in the sense that Anderson uses " Normal ". The latter term is widely used without any genetic implication for a situation where a fault dips towards the downthrow side—usually but not necessarily a dip-slip fault.

(as in Text-figs. 2 and 4*a*) where the relative magnitudes of the principal stresses are co-ordinates. If we accept that one principal stress is approximately vertical, one corner of the triangle may be so designated. The question then arises whether the remaining two can be distinguished from each other; if not, only half the usual triangle is required (see Text-fig. 4*a*). However, in the usual picture of an orogeny, an initial compression is the first distinctive direction, and if this is regarded as the primary horizontal stress (*H*) the other may be described as lateral (*L*). It is then reasonable to employ a complete equilateral triangle with apices *H*, *V*, and *L*, and to distinguish the



TEXT-FIG. 4.

(*a*) Stress triangle showing three biaxial fields characteristic of gravity, wrench, and thrust faults. The second half of the triangle depends on a distinction between the horizontal stresses as primary and secondary.

(*b*) Stress triangle, assuming this distinction between primary horizontal stress (*H*), secondary horizontal stress (*L*), and vertical stress (*V*), shows distribution of (i) isometric, (ii) uniaxial negative, (iii) uniaxial positive, and (iv) biaxial conditions.

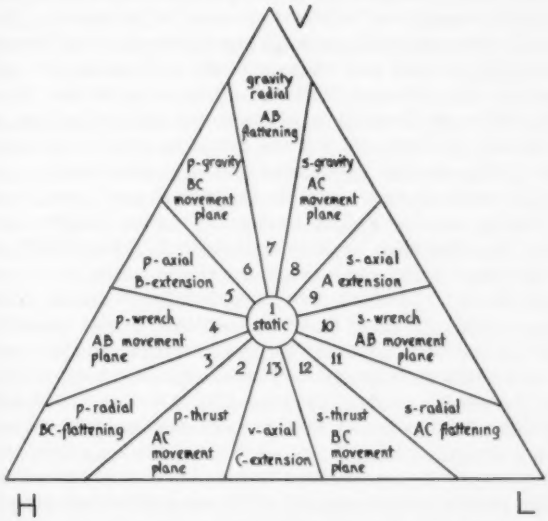
two halves $H > L$ and $H < L$ as primary and secondary. Conditions in the secondary half, though not so common, may be important and illuminating. In such a stress triangle the divisions in the primary half of Text-fig. 4*a* may be repeated in the secondary half (see Text-fig. 4*b*); there will, therefore, be, in the terminology of Table 1, six biaxial fields meeting in an isotropic point and separated by six limiting lines of uniaxial symmetry (three positive and three negative).

It has already been noted that to give the uniaxial conditions practical significance it is necessary to include cases of approximate equality: by applying the extension of usage again, the lines inside Text-fig. 4*b* are made finite fields. Thirteen separate regimes result (Text-fig. 5); it is the purpose of the following section to discover

how far these regimes may be recognized in nature and how far the concept is useful in tectonic analysis.

IV. Thirteen Tectonic Regimes

(1) The *static regime* is imagined as due to equality or near-equality of H, L, and V so that no movement results. It is the starting point and end point in any tectonic sequence, it may occur as a transient intermediary in the passage from one field to another, and it may



Text-fig. 5.—"Stress triangle" depicting thirteen tectonic regimes developed from Text-fig. 4 (*b*) and defined in section IV of text.

persist even during an orogenic episode at depths below the crustal disturbance.

(2) The *primary thrust regime*, in which $H > L > V$, is recognized by movement in the AC plane, with compression in A and extension in C[1] (Text-fig. 1*e* is a representation). The compression structures such as thrusts, folds, and cleavage commonly depicted in an AC tectonic profile belong to this regime, which may be regarded as the general condition throughout the development of an orogeny until relaxation sets in.

[1] Throughout the remainder of this discussion A, B, and C will be used as the regional axes, which may be taken as parallel to the principal stresses H, L, and V respectively.

(3) The *primary radial regime* follows from the condition $H > L = V$, where the primary compression is accommodated both upwards and sideways. This may be represented by some examples of slaty cleavages where, for instance, fossils have suffered flattening generally, not simply in the course of a linear extension.

(4) The *primary wrench regime*, in which $H > V > L$, is one of movement in the horizontal plane AB, with compression in A and extension in B, While extension is shared between B and C in the previous regime, B-extension is the dominant characteristic of this one and the next.

Some of the largest known strike slip fault systems may be difficult at this stage to relate to a total movement or stress picture until the movements of the crustal blocks in relation to each other is understood. The type of wrench fault clearly characteristic of this regime cuts a fold belt obliquely and the strike slip gives a net extension in B. In the familiar case of the Jura mountains faulting gives B extension which appears to be the result of overall curvature of the fold belt so allowing L to fall below V. In the straight fold belt recently described from Ny Friesland (Harland, 1956 and MS.) extension structures include boudinage and tectonic thinning as well as oblique faults. To allow B- extension without overall curvature requires differences along the strike such that V may exceed L for some distance. The conditions giving rise to this situation—relevant also to the next two regimes—will be considered in Section V below.

(5) The *primary axial regime*, defined by $H = V > L$, shows effects of vertical compression as well as horizontal compression accompanying elongation along B. Sander's term *Einengung* refers to this axial symmetry of movement (cf. Text-fig. 1*b* and *c*) but interest has hitherto centred on the symmetry of the associated fabric rather than its tectonic orientation. This fabric if lineated in A would represent regime 9, but when in B it represents this regime and indeed it is thought likely that most cases of this texture described in isolation (e.g. Weiss, 1954) belong to this regime. This type of movement may have taken place in the Ny Friesland axial belt at least as a transitional condition between 4 and 6.

(6) The *primary gravity regime*, in which $V > H > L$, is recognized by movement in the BC plane, with compression in C and extension in B. Among Anderson's "normal faults", normal dip-slip cross-faults in a folded region show this regime. Any B- extension in the main axial belt will encourage lateral relaxation above the level of maximum compression and so allow this condition at the surface.

(7) The *gravity radial regime* follows from the condition $V > H = L$ in which vertical compression is accommodated by radial movement outward in a horizontal plane. Shield volcanoes are an example of

this regime apart from an orogenic environment. Within a fold belt this condition will obtain between regimes 6 and 8 and may show a combination of effects typical of each. Thus, flat-lying gravity folds (see (8) below) may combine at the surface with gravity cross faults or at greater depth with axial plane cleavage admitting axial extension.

(8) The *secondary gravity regime*, with $V > L > H$, is principally a supra-crustal condition. Volcanic collapse structures, landslides, and mud flows are examples of the effect not necessarily related to crustal compression; but in the orogenic sequence where a distinction between H and L is important the necessary condition is where a welt has been built up by primary compression (H) above the level of the foreland. Structures resulting from superficial slope have been described extensively in the literature and summarized recently under the title "Gravitational gliding tectonics" by de Sitter (1954). The authors would not go so far as some, e.g. van Bemmelen (1954), in attributing most orogenic thrusting to gravity conditions. However, the development of large nappes or slides may require a secondary gravity regime throughout much of their extent while originating in a primary thrust regime at their roots. In a pile of nappes any extension in A, however produced, will cause a fall in H in neighbouring parts and possibly cause local tension in H. Movement within the fabric to express this regime might include Anderson's (1948) "Canal flow" in A.

Extension can, however, only take place in the direction of the primary compression at another level than that compression. This may be due either to the development of a surface slope, or to large-scale over-thrusting with local tectonic couples and possibly even tension in H.

(9) The *secondary axial regime*, in which $L = V > H$, again has movement opposed in direction to the primary compression. It will, therefore, occur only above foreland level or else after the orogenic climax has passed and relaxation has begun. Ideally this regime would be expressed in flow possibly related to "canal flow", but at the high level envisaged folding and related structures combining features of regimes 8 and 10 are more likely.

(10) The *secondary wrench regime*, defined by $L > V > H$, has essentially planar movement (in AB) with compression in B and extension in A. Cross-folding on steep axes will give this effect (see Text-fig. 6). Folds of this orientation in metamorphic rocks have long been known as *Schlingenbau* in the Alps, and more recently described by Dr. Sutton and his co-workers in the Scottish Highlands (e.g. Ramsay, 1957) as steep axial plunge secondary folds.

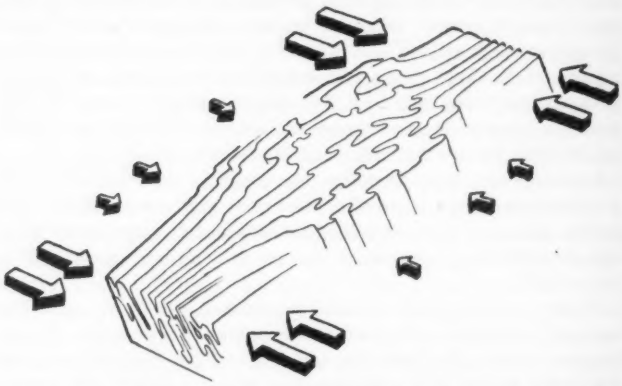
The secondary compression along B necessary in regimes 9 to 12 can arise only in a situation adjacent to one in which primary compression in A causes extension along B. Indeed, neither extension nor

compression in B are likely to obtain generally except in relation to each other, as will be shown below (section V).

(11) The *secondary radial regime* with $L > H = V$ has radial extension or flattening in the AC plane. Cleavage in AC related to cross folding (regimes 10 and 12) might reflect such a regime. The conjugate fold system in the Moine Thrust zone described by Johnson (1956) with fold axes parallel to the thrust motion in A and elongation in that direction may illustrate this regime as a local phenomenon and also regime No. 12.

(12) The *secondary thrust regime* is defined by $L > H > V$ with compression in B and extension in C (i.e. movement in the BC plane). Reversed dip slip cross faults could express this regime but it is better known from cross folds with flat-lying axes.

Cross folding on a large scale has long been known from many orogenic belts (e.g. culminations and depressions as in the Alps).



TEXT-FIG. 6.—Diagram showing local exaggerated development of secondary wrench regime with steep plunge axial folding adjoining two regions of B- extension.

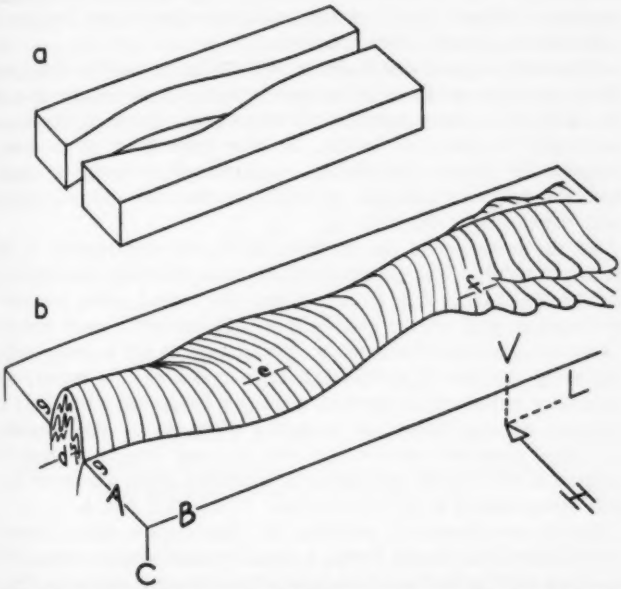
More attention has recently been given to small-scale cross folding and some of the literature has been summarized by Rast and Platt (1956) and still more recently by Haller (1957).

The generation of flat cross folds in this regime, as of steep cross folds in regime 10, may be due to secondary compression in B following a primary compression in A. It may be difficult to distinguish such a sequence of folding from periclinal or more complex structures belonging to regime 13.

(13) The *vertical axial regime*, with $H = L > V$, has vertical up- (or down-) squeezing promoted by radial compression, i.e. along both

A and *B*. Salt domes, volcanic vents, and plutons illustrate its symmetrical nature. Movement is facilitated by density and plasticity contrasts and is frequently far removed from obvious compression structures.

Within orogenic belts periclinal or more complex fold patterns may reflect conditions with secondary horizontal as well as primary horizontal compression. Migmatitic development, often associated



TEXT-FIG. 7.—Model of orogenic welt illustrating section IV of text. (*a*) The "jaws" shown without the welt which is partially depicted in (*b*).

with such complex folding, may represent a deep condition of this regime. In the Pennine Alps and in the Caledonian zone of East Greenland (e.g. Haller, 1955) folding is locally so complex that it is more probably the effect of one radial compression than a sequence of simple linear compressions.

V. An Orogenic Sequence of Regimes

In the foregoing enumeration and illustration of regimes, close inter-relation has become apparent. At a moment when in neighbouring parts of an orogenic belt different biaxial regimes are in

force, uniaxial conditions are to be expected at the boundary between them. Moreover, any of the uniaxial regimes may be regarded as the simultaneous combination of two biaxial regimes—an effect which may be difficult to distinguish from that of two biaxial regimes in succession.

These relationships, while making for difficulties in the interpretation of any particular situation, nevertheless encourage the consideration of their overall relationship in the development of a typical orogeny. Indeed, many of the regimes proposed can be shown possible in a simple model as follows.

The situation imagined is shown in Text-fig. 7*a* and *b*. The forelands are supposed to be of irregular shape so that compression of the axial belt is more severe at (*d*) than at (*e*). To show the shape envisaged, the jaws in Text-fig. 7*a* have been given finite depth, whereas for realism they should be considered to merge at depth with the deformed material. *H* is defined when the orogeny begins with the closing of the jaws.

At the inception of the orogeny, (*d*) is the first locality to be squeezed tightly. The compression promotes thrusting and folding; regime No. 2 ($H > L > V$) is in force. As vertical piling proceeds, the vertical load on (*d*) will increase and upward escape will be hindered. Since at this time the compression at (*e*) is quite small, squeezing sideways is an alternative; as soon as this occurs, the movement pattern at (*d*) becomes somewhat radial and field No. 3 is entered; vertical piling and streaming along *B* are simultaneous. It is now proposed that the pile over (*d*) may become so high in contrast to the vacancy at (*e*) that up-squeezing ceases altogether and only lateral escape from (*d*) continues. This is field No. 4.

Let us now observe the effects at (*e*). Supposing a second intense compression at (*f*), (*e*) forms a node toward which material is squeezed from both sides. (*e*) has hitherto suffered less intense effects of *H*, i.e. has been in a mild No. 2 regime, with a little vertical piling in progress. The vigorous incursion of material from either side will supplement this piling; now movements inward along *A* and *B*, together with extension up *C*, show the *V* axial regime, No. 13.

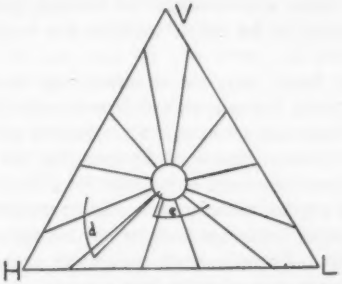
As at (*d*), the growing vertical pile will increasingly hinder vertical escape. Lateral in-surge may cease before the primary motion is halted or with a different degree of contrast between (*d*) and (*e*), insurge along *A* may be stopped while movement continues along *B*. This is regime No. 12 where cross-folding on flat axes may be found. It is also proposed that the vertical pile raised by the *B*- surge may be such and the compression at (*d*) so great that extension in *A* begins at (*e*) by retreat of the jaws (or by loss upwards of material near them). The accommodation of the *B*- surge along both *A* and *C* together

constitutes the radial regime No. 11. Accommodation along *A* alone will give regime No. 10 with steep axial cross folds.

In the supra-crustal pile the occurrence of gravity slides at localities such as (*g*) would show regime No. 8. Although it is easy to imagine the vertical part of the movement as downward from above, upward movement from below is equally important; comparison with a breaking wave is justified. In fact, it may be said that vertical descent does not occur except by relaxation.

It may be noted that even if regime No. 8 appears at the surface before regime No. 4 is established below, the latter may still develop; the height of the vertical pile is not limited to a ceiling by the realization of regime No. 4, although the rate of growth is greatly reduced.

Relaxation effects are capable of reversing all the movements and



TEXT-FIG. 8.—Stress triangle showing two sequences of development in localities (*d*) and (*e*) of Text-fig. 7 (*b*).

thus giving the regimes in the upper part of Text-fig. 5. Besides events in the orogenic decline, small relaxations arise through the rigidity of real rocks: for rigidity permits an equilibrium position to be passed and on eventual rupture the excess movements are reversed. In neither situation, however, can relaxation events match the scale of the progressive movements.

VI. DISCUSSION

Whether or not examples of these regimes have been correctly placed the concept stands as an analytical criterion in tectonic interpretation. It may be used even though there are many practical difficulties which limit its use and some theoretical implications which may need later clarification. Some of these points will be reviewed below.

The overriding difficulty—both theoretical and practical—is the relation between stress and strain. These do not always coincide in

their principal directions or even in their relative magnitudes in corresponding directions (allowing that one would be the inverse of its counterpart). One reason for the complex relationship is the mechanical inhomogeneity of the earth's crust, and another is that a pattern of relatively small stresses may fail to make a similar pattern in the deformation of relatively strong material. Both limitations may be removed on a sufficiently large scale. This difficulty cannot, however, be entirely eliminated by " bulking " since a scale sufficiently great to resolve the theoretical difficulties often exceeds the total bulk of material available or belonging to one regime.

It follows from this that the resulting structure is more likely to reflect the appropriate regime in plastic flow (than in rigid deformation)—or in other words, a flow movement is capable of expressing the regime in a smaller unit (e.g. a hand specimen) whereas heterogeneous deformation with mixtures of faulting, folding, and shear movements require to be taken together (or bulked) to find the overall effect.

On the other hand, any one structure may be the result of a sequence of regimes, but a rock fabric may only indicate the most intense or the latest one, whereas if all structures are taken into consideration there is more chance of disentangling the whole sequence.

The application of the same stress field for different durations may lead to different degrees of deformation and conversely a static stress field will be changed instantaneously by any minute strain. However, by attempting to infer the overall movement some relation to the effective stress pattern may be obtained.

Structural symmetry, and the relation between geometric, kinematic, and dynamic patterns has been the subject of much study, particularly in structural petrology. Structural symmetry (which includes consideration of the elements of primary and secondary foliation and lineation) and bulk symmetry are mutually dependent but not identical because structural symmetry concerns the local effect which may be different from the bulk effect and may include structural elements (e.g. the effects of rotation and of stress on the growth of minerals) which may not be directly translated in terms of bulk movement. In a recent review of the field Turner (1957) has shown the need to take more account of the bulk movement pattern. In general, the latter will have a higher symmetry than the separate constituent movements and structures.

Another source of difficulty arises from the general assumption that the principal stresses will be approximately vertical and horizontal. This orientation may be rotated in boundary situations between distinct media or units when some tectonic couple is exerted. Such departures from vertical equally qualify the categories of thrust,

wrench, and gravity faults which have nevertheless proved to be of general use.

A regime may in theory be specified at any one time and place by a point on the stress triangle and for the foregoing reasons precision is liable to be far less. With this in mind thirteen fields, to be used singly or in groups, were thought to be sufficient for definition. While the arrangement on the triangular diagram gives a continuously connected series of situations it is quite arbitrary as to how the triangle is divided into fields and the conventions chosen may not be the best. It may be pointed out that biaxial positive and negative fields would divide the triangle into six of each separated by lines. In preference to this a practical regrouping of the situation has been adopted, in effect taking each of the rock boundary conditions as typical of a field so grouping extreme positive (elongation) by extreme negative (flattening) movements separated by mid " biaxial " (predominantly planar movement). The names have been extended by analogy from other uses and no new terms are proposed. If the present use of tectonic regimes be accepted it will, in combination with the other terms used, save further jargon.

The concept of a tectonic regime involves symmetry (of bulk strain) and orientation of strain; that of tectonic facies (Harland, 1956) involves symmetry (of bulk strain) but also involves particular forms and scales with their distinctive compositions within one overall deformation. Strain is thus involved in both the facies and regime concepts but whereas the bulk pattern of strain (and implied stress) is common to both, orientation is peculiar to regime, while the detailed structures due to the individual properties of the rocks reflect more fully the conditions of formation and are peculiar to facies. A variety of facies may be reflected in structures belonging to one regime so that the number of tectonic styles possible for a single movement pattern is considerable and even greater for a sequence of facies and regimes.

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The Development of the Mid-Zambezi Valley in Northern Rhodesia since Early Karroo Times

By R. Tavener-Smith

Abstract

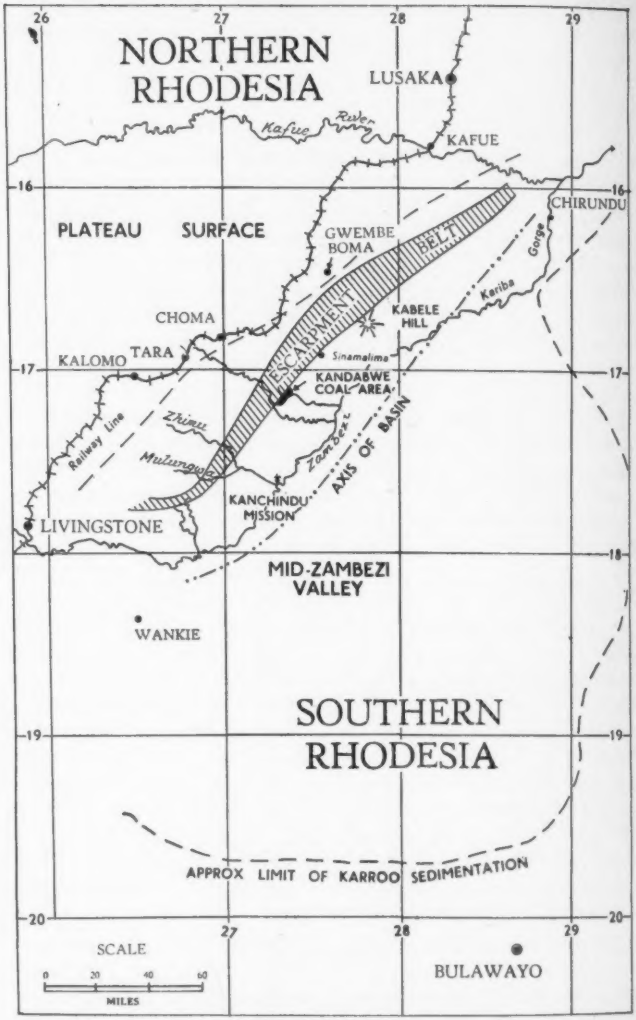
Karroo sedimentation in the mid-Zambezi area presents an instance of deposition in an intra-cratonic basin during a complete climatic semi-cycle. Mention is made of the tectonic factors which controlled the course of Karroo sedimentation in the area, and the nature of the pre-Karroo surface is commented upon. An account is given of the depositional history of the mid-Zambezi basin during the Karroo period. The structural development of the region since Karroo times is outlined, and it is suggested that the present valley is essentially a downwarped and eroded part of the peneplain which still forms the adjacent plateau to the north-west.

Introduction

Location.—The depression in which the Zambezi River has its course between the town of Livingstone and its confluence with the Kafue is known as the mid-Zambezi valley. The writer has recently spent three field seasons in mapping that part of the valley which lies between Kabele Hill and Kanchindu Mission, on the northern side of the river.

Physiography.—A large part of Northern Rhodesia is a high plateau standing at about 4,000 feet above sea level. In the mid-Zambezi region this surface is represented by the Lusaka–Kalomo upland which falls gently towards the Kafue basin on its western side, but is terminated more abruptly by the so-called escarpment overlooking the Zambezi River to the south-east. Except in a few places the latter feature is neither a structural nor an erosional scarp in the strict sense of the term, but is a belt of deeply dissected country from 2 to 15 miles wide, which probably represents the downwarped and eroded edge of the plateau. The floor of the mid-Zambezi valley stands at about 2,000 feet above sea level and has an undulating relief. The descent to the river from the Salisbury–Bulawayo ridge is much more gradual than is the case on the north-west side of the valley, and there is no deeply dissected zone comparable to the escarpment belt.

Regional geology.—South-west of Gwembe Boma the escarpment country and adjacent plateau are largely underlain by crystalline metasediments of the Basement Complex, and associated granitoid rocks of the Choma-Kalomo batholith. Recrystallized limestone and schist of the Katanga System (late Pre-Cambrian) outcrop extensively over the corresponding areas to the north-east. Karroo rocks are largely confined to the floor of the mid-Zambezi valley where they owe



TEXT-FIG. 1.—Index map to localities mentioned in the text. The Salisbury–Bulawayo ridge extends north-eastwards from Bulawayo for about 200 miles.

their preservation to a combination of downwarping and faulting. They comprise a maximum of over 10,000 feet of continental sediments ranging in age from late Carboniferous to early Jurassic, and rest with marked unconformity upon the metamorphosed Pre-Cambrian rocks below. A few small outliers, mainly of Upper Karroo age, are present in the escarpment belt and on the adjacent plateau as, for example, near Tara. They mark the former extension, at certain times, of the sediments from the basin onto the adjacent positive structural features.

Structure.—The flank and floor of the mid-Zambezi valley are extensively faulted on the Northern Rhodesia side, where the net result of the movements has in many cases been the formation of a number of tilted blocks sloping towards the river, between the edge of the plateau and the Zambezi (Text-fig. 2, No. 6). The tilted blocks are bounded in each case by normal step faults hading towards the plateau, and the latter are usually antithetical to major boundary faults that delimit the mid-Zambezi valley on its north-west side. The boundary faults in most instances throw rocks of the Karroo System against those of the Basement Complex, and trend roughly parallel to the course of the Zambezi River, towards which they hade steeply. In some places, however, where master faults are apparently absent, the step faults appear to be a function of downwarping movements which had their maximum expression in what is now the escarpment belt. Even so it is possible that in such cases major boundary faults do exist, but that their presence is not made manifest as they affect only rocks of the Basement Complex.

It seems likely that the localization of warping and faulting along the north-western side of the basin is due to the presence of an ancient line of crustal weakness separating the structurally positive area of the Choma-Kalomo batholith from the structurally negative mid-Zambezi valley. Relatively rapid subsidence in the vicinity of this linear zone of weakness during Karroo times led to the deposition of the greatest thickness of sediment there. The sedimentary axis of the basin is therefore situated much nearer to the north-west than to the south-east side of the basin, which is thus asymmetrical in cross-section.

Geological History

General

The mid-Zambezi valley was a structurally negative area long before the commencement of Karroo times (Du Toit, 1954, pp. 204, 559) and the preservation of Karroo rocks within it demonstrates that it is so still. The present valley occupies the site of one of a number of intra-cratonic basins, which existed on the surface of the African shield

during the Karroo period. The basin was formed, and subsequently extended, as a result of gentle epeirogenic movements affecting the continental mass, which also caused the uplift of neighbouring positive areas. The erosion of the latter provided the sediment which entered the basin until it was filled and, with the wearing down of relief on the adjacent positive areas, a state of equilibrium was attained. Throughout Karroo times deposition more or less kept pace with the rate of subsidence in the basin, and similarly on the positive areas erosion varied in intensity with the amount of elevation. It is unlikely that at any time a deep physiographic valley occupied the site of the basin, or that excessive relief occurred on the positive areas.

Properly speaking, the Karroo sediments in the mid-Zambezi area

EUROPEAN EQUIVALENT			SERIES	SYMBOL	FORMATION	THICKNESS
KARROO SYSTEM	UPPER	LOWER JURASSIC	STORMBERG		BATOKA BASALT	NOT KNOWN
		TRIASSIC		K7	UPPER KARROO SANDSTONE	6,300ft +
				K6	ESCARPMENT GRIT	ca.1,250ft
	LOWER	PERMIAN	LOWER BEAUFORT	K5b-e	MADUMABISA MUDSTONE	1,500-2,150ft
			ECCA	K2-K5a	GWEMBE COAL FORMATION	100-550ft
		UPPER CARBONIFEROUS	DWYKA-ECCA	KO-K1	LOWER WANKIE SANDSTONE	0-250ft
UNCONFORMITY						
SCHISTS AND GNEISSES OF BASEMENT COMPLEX						

TABLE 1.—Geological formations present on the Northern Rhodesia side of the mid-Zambezi valley.

present a record of the filling up of two distinct basins, which existed successively on approximately the same site within a short space of one another. The Lower Karroo rocks represent the sediments of the older basin which was initiated early in post-glacial times, and was filled by the end of the Lower Beaufort. The second was brought into being by the increased tectonic activity which was a prelude to Stormberg deposition, and eventually became filled with the sandstones and basalts of the Upper Karroo. Each of the two cycles commenced with rudaceous and coarsely arenaceous deposits, and concluded with sediments of much finer grain. They are separated in time by the hiatus in sedimentation which occurred during the Middle and Upper Beaufort, and differ from one another in the character of their component strata. The differences are best understood with reference to the climatic changes which occurred during the Middle and Upper

Beaufort, and the contrasting environmental factors which governed the derivation and deposition of the sediment in each case. It is not proposed to discuss environments at length here, but as regards climatic conditions it may be said that Karroo deposition took place during a complete climatic semi-cycle. The lowest beds are the result of glacial activity, whilst the uppermost were the outcome of large-scale vulcanicity in an arid desert climate. The intermediate strata represent a progression from one extreme to the other. The fine-grained sediments of the Lower Karroo cycle are of lacustrine and paludal origin, and were laid down under cold to cool humid climatic conditions. Those of the Upper Karroo are dominantly arenaceous and of terrestrial origin, being deposited in a semi-arid or arid climate.

Nature of the pre-Karroo surface

The pre-Karroo land surface in the mid-Zambezi valley was mature in character. No instance of an " island " of old metamorphic rocks protruding through the lowest Karroo sediments has so far been encountered, and the inliers of older rocks at present to be seen in the valley owe their presence, without exception, to faulting. The thickness of the Lower Wankie Sandstone, which varies up to 250 feet, was apparently sufficient to blanket the relief on the pre-Karroo floor, as no instance is recorded of rocks of the Gwembe Coal Formation resting directly upon those of the Basement Complex. In the Kandabwe area, where the basal part of the Lower Wankie Sandstone is composed of glacial and fluvioglacial sediments, variations in the thickness of those beds suggest the presence of relief of about 150 feet on the pre-Karroo floor. During the glaciation continental ice must have exerted a subduing effect upon the topography, and it seems likely that, following the withdrawal of the ice, deposition of glacial detritus in hollows where it would be preserved from subsequent erosion may have gone far towards the elimination of pre-Karroo relief.

The regularity of the pattern of the Basement-Karroo unconformity in relation to present topography tends to confirm that the pre-Karroo surface in the region was mature in character. Bold relief, if present, would be reflected in the pattern of the unconformity.

The findings of the writer in this matter are in accord with those of H. S. Gair, who mapped the adjacent north-east part of the mid-Zambezi valley (Gair, in press). The pre-Karroo floor beneath the extensive Witbank coalfield in the northern Transvaal has also been shown to be of mature relief (Wybergh, 1922, p. 40). It has been said of the pre-Karroo land surface north of the Limpopo-Zambezi watershed that, " There is good evidence to believe that . . . the Karroo beds were deposited as a sequel to a prolonged period of denudation so complete that only rarely are there any relics of older

sedimentary formations between the Karroo strata and the crystalline rocks beneath " (Mennell, 1929, p. 264). It is logical to infer from this that the writer considered the pre-Karroo surface to have been mature in character.

An interesting contrast to the above conclusions is afforded by the work of Dixey and others, who have recorded features of strong relief on the pre-Karroo floor, and have suggested that the surface was, in general, one of considerable irregularity (e.g. Dixey, 1937*b*, pp. 80–5). The present topography of central and southern Africa shows, however, that the presence of local features of strong relief is not anomalous upon a wider surface of mature character.

Glaciation and post-glacial deposition

The Dwyka ice-caps are said to have formed over several centres, from the Windhoek area to northern Transvaal (Du Toit, 1954, p. 276), and the ice is thought to have extended between latitude 2 degrees north and the Cape of Good Hope. Veatch, however, thinks that the ice centre may at one stage at least have been further north, and names the area of the Luano valley in Northern Rhodesia (Veatch, 1935, p. 150).

About 900 feet of glacial conglomerate is present in the Lower Lukuga formation of the Belgian Congo, and over 1,000 feet of tillite is recorded from many places in the Cape Province of South Africa. In the Rhodesias deposits of Dwyka age are thin and patchy in distribution. This is in keeping with the theory that near the centre of a continental ice sheet ground moraine is likely to be less plentiful than near the margins.

As the ice retreated a residue of tillite and fluvio-glacial material was deposited. The coarse, sandy matrix of the fluvio-glacials and the degree of rounding of many of the pebbles points to the presence of much seasonal melt-water. Varved sediments and mudstone containing ice-rafted boulders were laid down in fresh-water lakes as, for example, in the Kandabwe vicinity (Tavener-Smith, 1955, pp. 19–22). Erosion following the deposition of the tillite and fluvio-glacial pebble beds resulted in their removal from all exposed positions, hence their patchy distribution to-day. Corresponding episodes of erosion are recorded from the Congo basin and the Union of South Africa.

Following the withdrawal of the ice, deposition continued in lakes occupying broad downwarps on the pre-Karroo floor, and the Lower Wankie Sandstone was laid down. Where best developed this formation consists of an upper division of sandstone, fairly constant in thickness, and a lower conglomeratic member, partly of glacial origin, and much more variable in its development. Much of the detritus was locally derived and rapidly " dumped " into the lakes, never being subsequently

reworked. A large proportion of it came, in the early stages, from the erosion of tillites and fluvio-glacial pebble beds. As the cold, wet climate which followed the withdrawal of the ice gradually grew milder, plant life became established, so that the upper part of the Lower Wankie Sandstone frequently contains carbonaceous streaks and even seams of poor coal.

At first the floor of the basin continued to subside as the sediment entered it, but, later, movement almost ceased, with the result that the low-lying adjacent land area could no longer contribute coarse-grained sediment. The basin slowly continued to fill until a state of near equilibrium was attained.

The Gwembe Coal Formation

The coal measures are composed largely of carbonaceous mudstone with bands of sideritic mudstone and occasional thin fireclays, which do not, however, underlie coal seams. Sandstone and shale are rare, and the coals are usually numerous, thin and of poor quality. The nature of the strata indicates that during this period the greater part of the mid-Zambezi basin was occupied by wide stretches of nearly flat, swampy country, thickly covered with vegetation, and alternating with extensive shallow lakes and semi-stagnant pools. The presence of the coal measures wherever rocks of Lower Karroo age are exposed in the valley shows that the coal swamps and lakes occupied the surface of the whole basin, and were not restricted to a number of isolated centres of deposition. A few scattered localities are known, however, where no coal seams occur, and the coal measures consist entirely of carbonaceous mudstone with subsidiary siltstone and sideritic mudstone bands, or concretions. Such areas are thought to have been the sites of shallow lakes during the greater part of coal measure time.

Conditions of deposition affecting the greater part of the basin were variable, since more or less continuous, small-scale heaving and settling movements resulted in the frequent interchange of areas occupied by lake and swamp respectively. Peat formation was constantly brought to a close over wide areas by relatively rapid subsidence and flooding, which caused the deposition of peaty clay and silt lake-bottom deposits until the lake was rendered shallow enough for vegetation to gain a renewed hold. The formation of peat then recommenced and continued until terminated once more by further subsidence. The tectonic instability of the period in most cases prevented the formation of thick coals, and resulted in the continual splitting of seams, and the occurrence of lateral transitions from coal into coaly, or carbonaceous mudstone. Lateral variations in the quality and thickness of coals over relatively short distances and the

difficulty experienced in correlating the seams of neighbouring localities may be attributed to the same cause.

Thick seams are present in a few places, such as Wankie in Southern Rhodesia, and the Kandabwe coal area and Mulungwa River in Northern Rhodesia. In all three cases the coal rests immediately on top of the Lower Wankie Sandstone. The seams vary in thickness up to 42 feet, 25 feet, and 14 feet respectively, and in each case grade both laterally and vertically into carbonaceous mudstone containing thin coals of no value. These thick seams, which are essentially local phenomena, must have formed under drier bog conditions in relatively stable areas which were, perhaps, at the time near the edge of the coal-forming region (Tavener-Smith, in press). In such a situation there would be a greater degree of stability than further into the basin and, in the presence of a suitable supply of vegetal debris, a thick seam could form. Ultimately, judging by the pattern of the Wankie and Kandabwe deposits, as the size of the basin increased, the former marginal areas themselves became affected by the unstable conditions of the interior, so that peat formation was temporarily terminated by flooding and carbonaceous muds were deposited instead.

An interesting parallel in some respects with the Lower Karroo coal measures is afforded by the extensive phase of peat formation which took place during the Holocene period over the North Sea area. The peats were formed following the withdrawal of the Pleistocene ice over the southern half of what is at present the North Sea, but which was at that time a huge morass, nearly at sea level. The cool and moist climatic conditions at that time must have been very similar to those obtaining when the Karroo coal seams were formed.

The Madumabisa Mudstone lake

The end of the Ecca and the beginning of the Lower Beaufort was marked by further subsidence of the floor of the mid-Zambezi basin. This finally terminated the formation of peat, and gradually the nature of the sediment changed from peaty mud and silt to grey and green mud, as the depth of the water increased. The presence of occasional thin, intraformational, mud-pellet conglomerates indicates that the lake was never particularly deep, and that the sediment was now and then exposed to the effects of sub-aerial erosion.

A cessation of the downward movement of the floor in middle Madumabisa Mudstone times caused the lake gradually to silt up once more, until in some places the sediments projected above the water as wide mud and sand banks. In addition to the dominant grey mudstone, thin sandstone and muddy limestone bands were formed, and give to this part of the formation its characteristic flaggy appearance. Ripple marks and the occurrence of several beds of

mudstone conglomerate testify to the general shallowness of the water. By this time the climate had undergone considerable amelioration, and a limited fossil fauna of small fresh-water lamellibranchs (including *Kidodia coxi*, *Palaeonodonta* and *Palaeomutela*), ostracods and fish scales is represented in the muddy limestones of this part of the succession. Bituminous limestone bands rich in plant remains suggest that some of the islands of protruding sediment supported a luxuriant vegetation.

Further gentle subsidence led to the deposition of thick, featureless, grey and green muddy sediments until, towards the end of Lower Beaufort times, the increasing dryness of the climate eventually brought about the disappearance of the Madumabisa Mudstone lake. It has been suggested that this part of the formation may consist to some extent of accumulations of wind-blown dust comparable to loess (Bond, 1955, p. 76). The reddish colour of the upper 40 feet of mudstone is thought to be the result of oxidation caused by weathering of the beds during the Middle and Upper Beaufort.

The phase of epeirogenic movements which had resulted in the deposition of the Lower Karroo ceased early in Beaufort times, and a state of equilibrium was attained between the worn-down positive areas and the basin, which had become filled with detritus. This condition persisted until the end of the Beaufort period.

Stormberg sedimentation

In contrast to the lacustrine sediments of the lower part of the System, the terrestrial grits, sandstones, and conglomerates of the Upper Karroo constitute a red-bed facies indicative of a semi-arid to arid climate. The sediments in general are characterized by their coarse grain, red colour, and frequently felspathic nature, which suggest rapid erosion, deposition and burial in an oxidizing environment (Pettijohn, 1949, pp. 449–51).

The commencement of Stormberg time was marked by a renewal of tectonic activity, and extensive upwarping of the surrounding positive areas brought into being highlands from which abundant coarse, clastic detritus was derived. In the course of these movements the peripheral parts of the pre-existing basin were warped and eroded, and the debris incorporated into the basal Stormberg sediment. The rejuvenation of neighbouring positive areas initiated a phase of vigorous erosion, and led to the formation of a second and more extensive basin of deposition in the mid-Zambezi area.

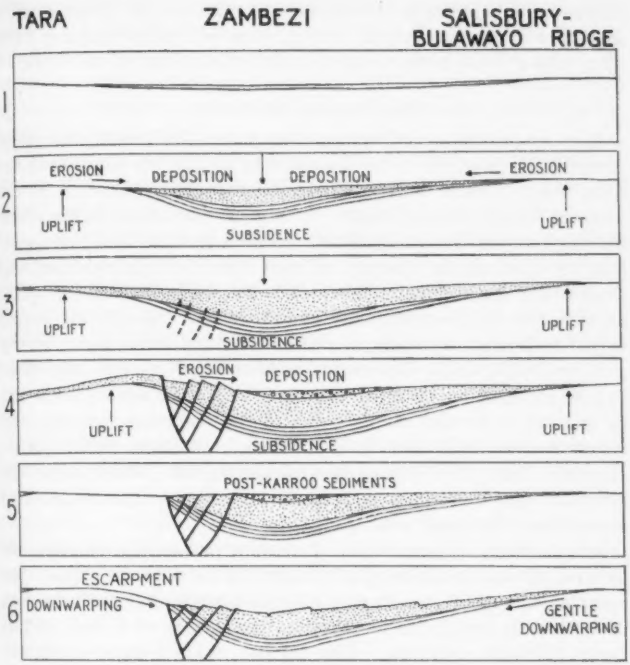
The up-arching movements attained their maximum expression at an early stage and diminished in intensity as Stormberg time progressed. With an abundant source of detritus at hand, the mid-Zambezi basin

soon became the receptacle for the thick deposits of rapidly accumulated, coarse, arkosic sediment known as the Escarpment Grit. The frequent occurrence of silicified wood in the lower part of this formation indicates that the climate, though becoming progressively warmer, was not yet arid. As the upwarping movements in the neighbouring source areas became less pronounced the supply of coarse material slowly ceased, and finer-grained sandstones with interbedded mudstone were laid down above the Escarpment Grit. In spite of the increasing aridity it is probable that some of the intercalated red mudstones were deposited in shallow playa lakes, and the occasional occurrence of thin bands of black carbonaceous mudstone points to the existence of a thick cover of vegetation locally. The thin Stormberg coals of South Africa are found at the same stratigraphic level in similar circumstances.

Transport of the Stormberg sediment seems to have been effected mainly by violent and short-lived stream activity and sheet floods caused by periodic rainstorms. Much of the Escarpment Grit is coarsely stratified and cross-bedded, suggesting deposition by such means. Thousands of tons of sediment per square mile are being moved in a comparable manner to-day after heavy storms in the Bushmanland part of South Africa (King, 1951, p. 90). The presence of thick deposits of aeolian sand near the top of the Upper Karroo sandstone indicates that towards the end of Stormberg time the wind assumed an increasingly important role as an agent of erosion and transportation.

During the Upper Karroo period sedimentation exceeded the limits of the earlier basin of deposition, overstepping the eroded edges of the warped Lower Karroo, and ultimately coming to rest upon rocks of pre-Karroo age. By the close of the period the basin was filled to overflowing, and the uppermost strata extended as a thin veneer over much of the eroded positive area which had previously been a source of sediment. As at the end of the Lower Beaufort the cessation of earth movements and consequent lack of sediment led to the establishment of a state of equilibrium.

The end of the Karroo period was marked by widespread eruptions of the fissure type, and a pile of basalt in places over 1,000 feet thick was accumulated (Gair, in press). The presence of beds of aeolian sand between the lower flows (the Nyamandhlovu Group of Southern Rhodesia) indicates that at first there was an interval of time between the eruptions. Later they must have been fairly continuous, as no interbedded sandstones exist in the upper part of the basalt. It is thought that by the end of Karroo time most of Southern Africa was an arid expanse of lava plains (Du Toit, 1954, p. 568). In the mid-Zambezi valley the Batoka Basalt is at present confined to the country in the Livingstone–Wankie area, and to more limited occurrences near



TEXT-FIG. 2.—Suggested stages in the development of the mid-Zambezi valley. (Diagrammatic and not to scale.)

(1) Commencement of deposition in broad downwarps on the pre-Karroo land surface.

(2) Pre-Stormberg movements caused the Lower Karroo rocks to be warped and eroded near to the margins of the basin. They were later overstepped by Upper Karroo sediments.

(3) Differential relations during Karroo times between the positive area centred on the Choma-Kalomo batholith, and the mid-Zambezi depression caused the old faults in the Basement Complex to be extended into the Karroo rocks.

(4) Extensive post-Karroo warping and faulting led to the formation of major boundary faults, with antithetical step faults hading towards them.

(5) During the subsequent peneplanation elevated areas were eroded, and the resultant sediment in part deposited in the mid-Zambezi depression.

(6) Downwarping with limited rejuvenation of faults. Subsequent erosion, with removal of the post-Karroo sediment, and formation of the dissected escarpment belt.

Chirundu and elsewhere. Dolerite dykes penetrate the Upper Karroo sandstone in the Kanchindu vicinity, however, which makes it seem likely that a much greater area was originally covered by the flows.

Post-Karroo developments

A pronounced phase of warping with associated faulting and tilting disturbed the sediments after the cessation of the late Karroo vulcanicity, and the present fault pattern largely dates from that time. It was followed by an extended period of tectonic quiescence during which the whole region was peneplained. In the course of this process the Karroo sediments were largely stripped from the higher ground, and some of the detritus was redeposited in the shallow depression which existed upon the site of the mid-Zambezi valley. The resulting surface is that to be seen at present at plateau level and forming the flat top of Kabele Hill, which rises several hundred feet from the valley floor to a height of about 3,000 feet above sea level. The surface at the top of this hill is bevelled across rocks of the Basement Complex and Upper Karroo which are faulted together. It seems likely that an ancestral form of the present drainage system existed upon the peneplain, and underwent some modification as a consequence of the movements which later took place.

Peneplanation was succeeded by extensive warping accompanied by a recrudescence of movement on some of the old fault planes. The peneplained surface was thereby lowered considerably over the area of the present mid-Zambezi valley, as the summit of Kabele Hill at about 3,000 feet indicates. This hill is a residual situated approximately 8 miles south-east of the main escarpment, and it is probable that further into the valley the peneplained surface was lowered even more. In the absence of more positive evidence as to the age of the last phase of downwarping, it does not seem unreasonable to connect it with the Tertiary movements which were of a similar nature elsewhere in Northern Rhodesia (Dixey, 1944, p. 34).

The subsidences led to a marked rejuvenation of the drainage system in the mid-Zambezi region. This was particularly the case along the north-western shoulder of the depression, where faulting and warping had been localized in a relatively narrow belt. In that area previously mature streams which had been consequent on the former peneplain became active agents of erosion once more and with their tributaries cut steeply downwards. The resulting dissection in a terrain consisting largely of soft schists ultimately resulted in the formation of the present Zambezi escarpment belt.

The same phase of erosion led to the removal of most of the post-Karroo sediment from the floor of the depression so that little of it now remains. It is possible that about 200 feet of poorly consolidated,

sandy conglomerate shown by drilling to rest upon Upper Karroo sandstone in the Sinamalima vicinity is the equivalent of the extensive pebbly sandstones and marls of Cretaceous age described by Dixey from the upper Luangwa valley (Dixey, 1937*a*, p. 69). Other occurrences of post-Karroo sediment have recently been found in the Kariba area (Hitchon, in press).

The drainage pattern over much of the Gwembe District suggests that the tributaries of the Zambezi were superposed on to the Karroo rocks from a previously existing cover of post-Karroo sediment; and is in marked contrast to the high degree of adjustment to structure shown by rivers in the escarpment belt. Further erosion of the Karroo rocks of the valley floor and the deposition locally of alluvium of mid-Pleistocene to Recent age have brought about the present configuration of the mid-Zambezi valley. Several stages in the evolutionary process as envisaged above are represented in Text-fig. 2.

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GEOLOGICAL SURVEY OF NORTHERN RHODESIA,
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Devonian Limestone Pebbles in Central Devon

By PHILIP F. HUTCHINS [1]

ABSTRACT

The presence of locally abundant pebbles of Middle Devonian Limestone in the New Red Sandstone near North Tawton, Devonshire, is recorded. It is suggested that these were derived in Permian times from a nearby outcrop of Devonian limestones which projected through the Culm Measures of the mid-Devon syncline.

THE Lower New Red Sandstone of Devonshire has been divided by Ussher (1906) into

Lower Sandstones
Breccio-conglomerates
Watcombe Clay

These beds rest with strong unconformity on the folded Devonian rocks and Culm Measures.

The outcrop of the Breccio-conglomerates extends northwards from Torquay past Dawlish and Exeter to beyond Tiverton. Near Exeter a narrow band of Breccio-conglomerate, following the Crediton valley, projects westwards from the main outcrop for about 21 miles.

The pebbles in these deposits have been the subject of an extensive study in the past (Worth, 1890 ; Ussher, 1902 ; Ussher, Reid *et al.*, 1913 ; Scrivenor, 1948). They consist mainly of fragments from the underlying Devonian and Culm rocks, but at certain horizons pebbles and boulders of igneous and metasomatized rocks are abundant. As might be expected, limestone pebbles are common near Teignmouth and Dawlish where the Breccio-conglomerates are near outcrops of Devonian limestones. Northwards away from the Devonian rocks limestone pebbles are less abundant.

THE CREDITON VALLEY

Limestone pebbles are rare in the New Red Sandstone of the Crediton valley. In the eastern outcrops very few have been found and calcareous cementation is absent. However, in the west, near North Tawton, quarries at Solland and Westacott are excavated in red sandstones and conglomerates which are heavily cemented by calcite. The conglomerates contain abundant limestone pebbles.

In this part of the Crediton valley the dip in the New Red Sandstone is predominantly southwards. The succession may be separated into

[1] This is one of a number of manuscripts which Dr. P. F. Hutchins left at the time of his death in 1955.

two main divisions. At the base the Cadbury Beds rest directly on the Culm Measures. The Cadbury Beds are characterized by pebbles of " Culm-type " sediments and vein-quartz only. The Bow Beds overlie the Cadbury Beds and overlap westwards on to the Culm Measures. These contain, in addition to pebbles of " Culm-type " rocks and vein-quartz, fragments of untourmalinized lavas. The Bow sediments are also interbedded with lava flows (the Exeter Traps). The horizon with limestone pebbles forms a sub-facies near the base of the Bow Beds.

A fuller account of the stratigraphy and petrology of the Crediton valley deposits will be published later.

Solland Quarry

At Solland, in the parish of Sampford Courtenay, there is a large disused and overgrown quarry on the northern margin of the outcrop of the Bow Beds (National Grid SS 615021). This shows a section of coarse limestone conglomerate resting with an irregular erosion surface on fine conglomerates cemented by calcite. The conglomerate contains pebbles of lavas and red-stained " Culm-type " sandstone and shale—the typical Bow Beds assemblage. The overlying conglomerate consists mainly of rounded and part-rounded pebbles of limestone. These have major axes varying up to an observed maximum of 25 cm. Small, rare pebbles of lava, vein-quartz and red-stained " Culm-type " sandstone and shale are also present. The exteriors of all the limestone pebbles are extensively corroded and much weathered. The calcareous cement of both the limestone conglomerate and the underlying beds is undoubtedly derived from the limestone pebbles by solution.

Most of the limestones are grey-white in colour with occasional pink streaks. They are massive and have suffered recrystallization and some shearing. Some are veined with coarse cream-coloured calcite. Rarer pebbles of a fine-grained black limestone are also present.

Bryozoa, stromatoporoids, brachiopods and corals have been found in the pebbles. Many are too altered for identification, but Dr. P. W. Taylor has kindly identified the following genera and species amongst the corals and stromatoporoids (cf Taylor, 1950):

Amphipora ramosa Phillips
Hooeiphyllum tenue Wedekind
H. cf. *normale* (Wdkd.)
H. cf. *gracile* (Wdkd.)
Grypophyllum sp.
Caunopora spp.

These specimens are in the Sedgwick Museum (H 5,010–H 5,020).

Dr. Taylor considers that the coral fauna indicates an Upper Givetian (Upper Middle Devonian) age for the limestone. The limestones have a similar lithology to those found *in situ* in the Chudleigh district of south-east Devon.

Solland Quarry is situated on the northern margin of the New Red Sandstone outcrop and abuts against the Culm Measures. On the north the outcrop is hidden by scree, but a small exposure of typical grey Culm sandstone with red streaks appears to be *in situ*. It is well above the outcrop of the New Red Sandstone in the quarry. This relative position and the almost unstained condition of the Culm sandstone suggests that the boundary may be faulted.

The presence of supposed Devonian limestone pebbles in the New Red Sandstone at Solland was first described by Vicary (1867). He found *Stromatopora* and thought that the limestone was similar to that found at Lummaton (Ussher, 1900, p. 57).

Westacott

Vicary (1867) also recorded a similar limestone conglomerate from Westacott, east of North Tawton and 4¾ miles east of Solland Quarry. This quarry (SS 685022) is now part-filled with water and most of the section is obscured by scree. A small exposure on the north side of the quarry shows hard, well-cemented calcareous sandstones interbedded with beds of conglomerate containing pebbles of limestone and "Culm-type" sandstones and shales. The limestone pebbles are similar to those occurring at Solland. Only one fossil—a coral—has been found. This was too altered for identification. As at Solland both grey-white and black limestones occur. The largest limestone pebble noted had a major axis 15 cm. in length. The beds dip 20°S with a strike of 185°.

Other Calcareous Rocks in the Crediton Valley

A further exposure of a calcareous facies occurs to the west in the bed of the River Okement, west of Exbourne, on the northern margin of the New Red outcrop (SS 5885.0260). Here calcareous sandstones dipping to the south overlie normal Bow conglomerates which are heavily cemented and veined with calcite. No limestone pebbles have been found but thin-sections of the calcareous sandstone show that detrital grains of limestone were originally present. The whole is now recrystallized, but rounded outlines of the grains are revealed by the original coating of haematite which may still be distinguished within the large plates of calcite. It may be noted that this exposure is on the same line of strike as the calcareous facies at Solland and Westacott.

Further south in the bed of the Rover Okement near Jacobstowe Ridge (SS 592018) the beds are slightly calcareous. South of the bridge a more calcareous conglomerate is exposed and one pebble of partly silicified limestone was found (SS 5924.0159).

Although this calcareous facies has not been detected between the

exposures at Westacott, Solland, and the River Okement, it is probable that it may form a definite horizon at the base of the Bow Beds.

East of Westacott calcareous rocks have not been found. However, in a boring for water at Sandford (approximately SS 829025) north of Crediton, a bed of " limestone ", 3 feet in thickness, is recorded at a depth of 248 feet. (The full record of the borehole is in the Geological Survey files, reference no. 325/6.) This " limestone " may be a calcareous sandstone. It is significant that the borehole is near the northern margin of the Bow Beds in that area. The " limestone " would therefore be almost on the boundary between the Bow and Cadbury Beds, i.e. at the same horizon as the calcareous facies in the west.

Vicary (1865) reports that the conglomerate at Randiscombe, West Sandford (? Ranscombe SS 7980.0305) was originally burnt for lime. If the locality is recognized correctly this is also on or near the boundary of the Bow and Cadbury Beds.

The occurrence of Devonian limestone pebbles in central Devon is surprising. No Devonian rocks are known from the region, the nearest outcrops being 16 miles north in the Exmoor region. However, the pebbles in the New Red Sandstone are of the southern facies of the Devonian, the nearest outcrops of which are near Chudleigh, 20 miles to the south-east beyond the eastern ridge of Dartmoor. The large size of the pebbles (major axes up to 25 cm. long) at Solland and their great abundance compared with other constituents all suggest a restricted local origin in the North Tawton–Hatherleigh area.

The New Red Sandstone outcrop is surrounded by Culm Measures. No outcrop of limestone is known from the area, although limestones of supposed Lower Culm age occur to the south on the northern border of Dartmoor and a bed of impure limestone was recorded near Hatherleigh by Sedgwick and Murchison (1840, Pl. 51, fig. 1).

The Culm Measures near Hatherleigh and North Tawton are supposed to be Middle Culm (Ussher, 1906), but Ussher (1892, p. 137), when recording the finding, by Vicary, of goniatites in the Culm Measures of Cocktree Moor, south of North Tawton, considered that these beds must be Lower Culm brought up by an anticline. Perhaps Devonian limestones were similarly exposed during Lower New Red Sandstone times.

In this connection it is worth noting that the Permian lava at Hannaborough, south of Hatherleigh (Tidmarsh, 1932), is very calcareous and may have assimilated some limestone before extrusion. To the south an " elvan " dyke (SS 528020) also contains abundant calcite. This suggests that limestones may underlie the area (these may, of course, be the limestones of the Lower Culm). The lava intercalated in the Bow Beds at Stone Quarry south of Westacott is also calcareous,

but it may have derived its carbonate from the calcareous Bow Beds which outcrop to the north and would, therefore, underlie the lava (the dip is to the south). In the south-east part of the quarry the lava rests on hard calcareous conglomerate containing a pebble of grey porcellanous limestone.

Conclusion

The abundant Upper Givetian limestone pebbles from Solland near North Tawton most probably had an origin in the North Tawton–Hatherleigh area. No outcrop of Devonian rocks is known from the area, but the Culm Measures are very imperfectly known and the limestone may yet be found. However, it seems more probable that the outcrop is at present hidden under the narrow strip of New Red Sandstone. Moreover, the absence of limestone pebbles in the higher part of the Bow Beds would accord with burial of the Devonian limestones under later deposits. The line of the Crediton valley was a zone of tectonic disturbance during early New Red time as shown by subsidence to accommodate the breccio-conglomerates and the igneous activity of the Exeter Traps and the " elvan " dykes south of Hatherleigh. It seems probable that Devonian limestones were also brought to the surface along this line.

Acknowledgments

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Alternating Pleistocene Morphogenetic Systems

By C. A. Cotton

(PLATE V)

Abstract

Alternation of morphogenetic systems conditioned by wide swinging of climatic belts equatorward and poleward has been used to explain otherwise puzzling relict forms in many regions. A characteristic finely dissected landscape relief prevalent in some middle-latitude regions is in strong contrast with smoothed, coarse-textured, whalebacked, relatively featureless relief in others, notably western Europe, which are now in the humid-temperate zone. This latter is believed to be relict from the periglacial regimes of Pleistocene ice ages. In New Zealand where the former landscape type is typically developed, there has also been quite strong "periglaciation", and in Europe running-water erosion cannot but have operated to some extent in interglacial ages. It appears that the periglacial smoothing has been pushed to the extreme limit in Europe but that its effects have been neutralized in New Zealand by the vigour of dissection in the alternating warm ages. Periglacial smothering, or "fossilization", of fine-textured relief, which has been followed by renewed fine dissection, is very common at Wellington, New Zealand. The reasons are not clear why running water should have been (and should still be) more efficient in dissecting the New Zealand landscape than that of Europe, so that the relict periglacial landscape remains intact in the latter region.

Morphogenetic Systems

An idea that has been attributed by Tricart (1951*a*) to A. Cholley is that of the "erosion system", the assemblage of processes or agencies locally or regionally operating under a prevailing climatic condition and combining to produce a particular type of landscape. For "erosion" system Tricart and Cailleux (1955, p. 41) now prefer to substitute *morphogenetic system*. This concept involves consideration in each instance of a complex of processes certain of which are shaping hillslopes and others river valleys and their profiles. These authors point out that it is not erosion alone that is involved but erosion associated with the production and deposit of "correlated sediments". Thus, just as the expression "geomorphic cycle" is preferable to "cycle of erosion", seeing that most landscapes comprise forms of accumulation as well as those due to sculpture by erosion, so "morphogenetic system" is preferable to "erosion system". It might be added that the former is better also because it does not exclude tectonic processes. Tricart and Cailleux (1955, p. 108), do not include in their specification discussion of the origin of initial forms, but concentrate instead on the erosive processes, with correlated processes of evacuation, transportation, and accumulation of the debris of erosion, emphasizing always the

climatic factor and always stressing the importance of a major boundary between non-forested and forested regions, in the latter of which they claim that a " sufficiently dense cover of vegetation " interferes very significantly with morphogenesis (p. 112).

Examples of morphogenetic systems, recognized and described by Tricart (1956*a*) in western Africa, are those of the equatorial rain forest (in Guinea), the savana, and the steppe. Boundaries between climatic zones having shifted quite considerably from time to time, in some cases forms that are now—slowly or rapidly—undergoing changes which will adapt them to one particular system (controlled by climate) are relict from some other system. Yet it is remarkable how some very peculiar assemblages of landforms, notably that of Fouta Djalon, Guinea, where no streams are graded, can be explained as due to control by processes unlike those of most other regions, especially regions with the " normal " regime, but peculiar to the climates now prevailing—climates that characterize the morphogenetic systems of these regions and differentiate them from others.

It would appear, however, that the systems of greatest interest to students of the landscapes that used to be classed together as " normal " —those, that is to say, of the better known and more habitable parts of the middle latitudes of the Northern Hemisphere and of Australia and New Zealand—are : (1) the humid-temperate rain-and-rivers ensemble as typified in New Zealand ; and (2) the periglacial, or cryergic, system, which has affected vast regions, especially in the Northern Hemisphere, in the Pleistocene ice ages, though superseded very recently in the middle latitudes by the rain-and-rivers ensemble which has been re-established.

Alternation

The very generally accepted doctrine of climatic oscillation, with four or more glacial ages having been separated by long interglacial intervals, makes it necessary to explain many landscapes as a result of the dominance alternately of one and the other of the rain-and-rivers and cryergic morphogenetic systems.

Some other attempts have been made to explain morphogenesis as due to development under more than one successive climatic regime characterized by radically different erosion processes, though not in every case in necessarily different climates. It is now usual to think of the most ancient and most extensive planations, which have truncated all rocks including the most resistant and of which traces remain as the classic " peneplains " and fossil " peneplains ", as products of pediplanation either in periods of widespread aridity and semi-aridity or under a similar morphogenetic system otherwise controlled—that is, independent to a great extent of temperature and rainfall but

conditioned by " biological deserts ", by the vegetationless state of ancient land surfaces (Tricart and Cailleux, 1952, p. 398; 1955, pp. 168–9; *also* Baulig, 1957, p. 924). Even the vast though less extensive peneplanations that affected Europe in Tertiary times were, it is now thought, climatically controlled : the temperature was consistently high and the climate must have been humid, with a corresponding great activity (according to Van t'Hoff's law) of chemical erosion and denudation, so that by a combination of downwearing of the interfluvial land surfaces with backwearing of the side slopes of already incised valleys rapid progress was made with destruction of relief, notwithstanding forest cover.

Alternation of contrasting climatic conditions has been invoked to explain the reliefs of those terrains with thick sandwiched formations of stratified rocks of different kinds which now form cuestas and escarpment-bounded structural plateaux. Humid periods of sapping at the bases of escarpments are believed by Mortensen (1947) to have alternated with others in which semi-arid pedimentation has been dominant.

Tricart (1951*b*) also has appealed to alternation of climates to explain the emergence as structural escarpments in the Paris Basin of different caprocks at different times. The underlying beds, wastage of outcrops of which has produced the escarpments, have offered, he claims, different resistance to ablation under the different (alternating) climatic conditions prevailing in the Pleistocene. The white Chalk, " normally " an escarpment-making caprock, became in cryergic episodes extremely weak, because it is particularly *gélive*. Being inherently frost-liable it has yielded very easily to ablation, thus exposing (increasing the relief of) escarpments of Eocene strata. Other strata, however, for the most part limestones, have emerged with prominent relief in non-glacial ages owing to the weakness—proneness to normal weathering and wastage—of Eocene and Jurassic strata exposed on their escarpment faces under the climatic conditions then prevailing. Tricart has diagnosed escarpment-making of both interglacial and preglacial (late-Pliocene arid) ages.

Contrast between Finely-dissected and " Subdued " Landscapes

A much more far-reaching application of the idea of alternation in Pleistocene times is required, however, to explain the contrast between the finely dissected landscapes of vast areas in lower middle latitudes (outside the more arid regions) and the very broadly subdued landscapes, with streams widely spaced, that appear in other (and in the Atlantic region rather higher) middle latitudes. It is not difference of latitude, however, that can be invoked to explain the profound difference in the general aspect of the relief ; for on the western and eastern

seaboards of the United States contrasting landscape types are found in similar latitudes. Nor can it be claimed that the west-European landscape, as it may be called, with its broadly subdued reliefs, is explicable simply as the typical product of the periglacial morphogenetic system operating for a long time without interruption, whereas the contrasting finely dissected landscapes of other regions are strictly " normal " in the sense of displaying the (unfinished) work of the rain-and-rivers system—that is to say, that they have altogether escaped the modifying effects of cryergic processes in Pleistocene ages. In New Zealand—notably at Wellington (Cotton and Te Punga, 1955*a*; Stevens, 1957)—periglacial cryergic activity has been intense.

When the position is examined from the point of view of what is known of Pleistocene history, moreover, the question as to what was happening in Europe in interglacial (interperiglacial) ages must be asked. As regards the possible or probable origins of these contrasting reliefs there can be little doubt that other finely dissected landscapes have been shaped in much the same way as the hills around Wellington, namely under forest and on terrains either of fresh rocks easily eroded by running water or of bedrock that is potentially resistant but is decayed to a very considerable depth (as is the case on the deformed Mesozoic greywackes at Wellington). Such terrain must be postulated to explain vigorous corrasion by very numerous, and therefore necessarily in most cases small, streams of water such as flow in a climate that is mild and rather humid and generally without a marked dry season. These are the conditions under which (at Wellington, for example) rapid incision of numerous small valleys has taken place, dissecting the surface to maturity.

Such dissection is largely a question of the evolution of land slopes, the broader slopes of a major relief being in this case thoroughly dissected. In some regions the prevailing " major " slopes are those leading down into the valleys of large rivers that have been incised by the almost world-wide Pleistocene rejuvenations. In California and New Zealand the slopes are in many cases tectonic, being on the flanks of blocks and strongly arched and domed initial features produced by Pleistocene orogeny.

The forms of major relief in the vicinity of Wellington were produced by such deformation of a peneplain in early or middle Pleistocene times (Cotton, 1957). The feral (unsubdued) fine-textured dissection of the Coast and Transverse Ranges of California (*cf.* Davis, 1912, pl. 7) strikingly resembles that common throughout New Zealand and quite probably resulted from similar causes. Photographs of landscapes in Japan indicate similarity there also. Though semi-aridity now prevails in Southern California dense forests undoubtedly extended far south on the Pacific Slope with each Pleistocene equatorward

swing of the climatic-zone boundaries. Large parts of New Zealand were forested until wholesale destruction of the "bush" took place within the last century (a deforestation that preceded the grassing of the hills shown in Plate V (fig. 1), and there is evidence of the existence of a more extensive forest cover in earlier (prehistoric) times.

Consequent–Insequent Dissection

On a surface with strong tectonic irregularity like that produced at Wellington by the Pleistocene orogeny (Plate V, figs. 1, 2), numerous consequent streams are formed that run down initial declivities. A potentially domical form which is due to irregular upheaval and is shown at the right in Plate V (fig. 1) has been dissected by radially disposed consequent streams so that it has become pyramidal. These and similar consequent streams may not have been as long to begin with, but all such streams can extend headward, and a close net of drainage lines is further developed by the branching out and headward extension of insequent tributaries. The prevalence of insequent branching is due to cratonization of the folded greywacke terrain and the depth to which the rock is incipiently rotted by weathering; these combine to make development of subsequent features exceptional.

When these very numerous consequent and insequent streams have deepened their valleys or ravines by vertical corrasion to forms with V section the result has been dissection to a fine texture that is by no means abnormal, though it might perhaps appear so to an observer accustomed to English landscapes or those of western Europe and unaware of their cryergic modification. As long as the ridges and spurs retain the sharpness they assume when close, or mature, dissection is first attained, the dissection pattern can be described as feral (sharp-edged, serrate, unsubdued). Whether for exactly the same reason as at Wellington or not, feral summit forms are characteristic of the areas of stronger relief in many of the regions with prevailingly fine-textured erosional relief. Vigorous ablation (wastage) of the land surface being in progress, however, as the intersecting slopes of progressively deepened ravines are themselves worn lower (without necessarily any backwearing of the slopes to gentler declivity) some softening of form must supervene, with rounding of summits and ridge crests to produce small-scale subdued forms such as characterize some fine-textured landscapes (*cf.* Davis, 1912, pl. 5).

An interesting demonstration of the fineness of such a pattern of dissection is obtained by comparing a topographic map (of, say, one inch to the mile) with either the actual landscape or with photographs of it. Whereas on a map (on the same scale) of a district in western Europe one finds all the natural features shown, in New Zealand this is by no means the case. Only a generalization of the relief

is possible. Every small valley branches, every branch ravine branches again, and so on almost *ad infinitum*.

West-European Subdued Relief

The forms of the fine-textured (commonly feral) landscapes, not only of Wellington and adjacent parts of New Zealand but also of many other warm-humid forested and recently forested regions, contrast strongly with those characteristic of other very extensive regions in corresponding latitudes that have wide spacing of streams and a resulting coarse texture of dissection (*Gliederung*) giving thoroughly subdued relief. This is the west-European type, which is widely developed in middle latitudes of the Northern Hemisphere, especially in the Atlantic region, where it appears to be coextensive with periglaciation.

Southern England, for example, is described by Te Punga (1957, p. 410) as a " typical periglacial landscape ". According to his account of it:

> Pleistocene denudation took place for the most part during periglacial episodes, for then the whole surface of the landscape was subject to vigorous downwearing as a result of mass wasting, in which solifluxion was particularly active. . . . It seems probable that the effects of successive periglacial episodes were cumulative, each later episode emphasizing the landforms produced in earlier episodes; it is unlikely that interperiglacial erosion, seeing that it was restricted essentially to linear processes, was competent to obscure or obliterate earlier developed periglacial landscape form.
>
> The rapid wasting of the land surface during periglaciation, due to the transportation of enormous quantities of material to lower levels, has produced a landscape of subdued aspect characterized by slopes that are convex near the top and concave near the bottom. . . . Such slopes . . . . dominate the present landscape.

Tricart (1951*a*, pp. 4, 5) has also described cryergic development of slopes of the land surface as typified in the eastern part of the Paris Basin thus:

> The upper part of the slope retreats. . . . The steepness of slopes is reduced, cliffs crumble, and enormous quantities of material descend to lower levels. The landscape form is (not stabilized but) alive. . . . Slope forms (thus developed) being almost immune from attack by the feebler processes of postglacial erosion still dominate present-day landscapes. In places where gelivation and solifluxion operated freely slopes generally have a form convex at the top and concave at the bottom, with an intermediate straight slope. This form is found on limestones as well as on granites, marls, and alluvia with fine matrix. The upper part of the slope is a zone of wastage . . . ; the concave basal part is a slope of accumulation . . . ; on the intermediate part thin coulées of soliflual debris are seen furrowing the rock. Slopes of this form, extremely common in France, cannot be explained as developing in the existing climate.

Landscapes wholly or largely shaped by the cryergic processes cannot be described as normal ; but on the other hand the fine-textured

"rain-forest" landscape as developed by running-water dissection in New Zealand and elsewhere is "normal" by definition. A change (and such a change must have taken place) from relatively fine, as in the latter, to ultra-coarse texture of relief finds a credible explanation in cryergic smoothing, which, besides some cryergic corrasion, involves a smothering of any pre-existing relief, coarse or fine, with a smooth-surfaced sheet of geliflual debris (head), so that dissection must begin anew, if there is to be any dissection of the surface at all. A small-scale parallel to cryergic landscape smoothing might be imagined if one pictured an intricate bas-relief of bronze laid out horizontally and treated by some vandal with a blow-lamp which fused the metal superficially and blurred the relief. "Normal" geomorphogenic theory is, on the other hand, at a loss to explain, though it recognizes the fact, that small dissecting streams are at some stage eliminated from many landscapes—"die Gliederung verliert ihre Feinheit" (Davis, 1912, p. 186) as the form is assumed to evolve from early maturity to the presumably more advanced, or older, stage Davis called "subdued" (1912, p. 284).

Parallel Development

Striking as the difference is and fundamental as it may appear at first sight, it may be possible to explain both the west-European type and at least that variety of the rain-forest type represented at Wellington as produced during *alternation* of morphogenic systems proper to cryergic (periglacial) and intercryergic (non-periglacial, interglacial, or normal) ages of the Pleistocene.

The day is long past when it was the fashion to ascribe detailed landscape forms, i.e., any other than the broadest and most generalized features of the relief, to pre-Pleistocene morphogenesis. Assuming then a comparatively modern origin of the relief without further argument, it is obvious that in the intercryergic intervals of Pleistocene time, some of which are believed to have been very long, "normal" dissecting processes must have operated freely. In New Zealand such intercryergic morphogenesis has obviously been dominant despite abundant evidence at Wellington of Pleistocene gelivation and solifluxion. The latter processes have, however, dominated the situation in southern England and western Europe.

Despite the evidence that the cryergic processes were successful in "subduing" the landforms, the general course of morphogenetic events in Europe throughout Pleistocene time can have differed little from that in regions that are now much more finely dissected. Major slopes were developing as a result of continental emergence (or progressive lowering of ocean level) which stimulated all large rivers to re-excavate their valleys. The side slopes of newly deepened and

widened major valleys are just as liable to dissection by minor streams as are newly developed tectonic slopes due to deformation of a peneplain such as took place at Wellington at some time in the Pleistocene. Perhaps in preglacial and certainly in interglacial times there must have been the same tendency towards development of fine or at least relatively fine-textured dissection. Relief so formed must, however, have been obliterated by processes of the cryergic morphogenetic system in the cold ages. Possibly dissection was at a minimum because of an unexplained slow tempo of normal erosion in Europe in the warmer Pleistocene intervals such as has been reported by Cailleux (1948, 1950) at the present day. A contrast of this with the fast tempo of dissection by streams which is obvious to any observer in New Zealand is very striking. There is at any rate no record of extensive " fossilization " under head in France of ravined landscapes which might have developed in interglacial ages. Tricart (1956*b*, p. 148) indicates that evidence of such fossilization is at least very rare.

Differentiation

Evidence of intermediate and successive stages being obscure, one is forced to resort to speculation regarding the undoubtedly complicated Pleistocene history of landscape forms and especially of the contrast, or growing differentiation, between the west-European and what may for the present be called the New Zealand type of landscape. Each, it may be assumed, has experienced successive morphogenies in which cryergic slope-smoothing and non-cryergic (normal) slope-dissecting processes have alternately been shaping landforms, the smoothing processes emerging triumphant in one region and the dissecting process in the other.

Apart from the probability of slow tempo of dissection in Europe (unexplained) in the ages in which conditions favoured the work of running water, it is probable that the smoothing and smothering effects of cryergy were considerably greater there. To cite an example of its (sometimes underestimated) potency, Suchel (1954, pp. 113–14) has described how in the Hils district, western Germany, a structural escarpment 250 feet high was worn back fully half a mile by frost weathering, a vast apron of the debris so produced being spread at the foot of the escarpment by solifluxion.

Dissection and Redissection of Slopes

It may be assumed from the spatial distribution of the landscape type so produced that fine-textured running-water dissection of slopes that are underlain by hard rocks is favoured by warm-humid chemical weathering of the rocks, such as has rotted the greywacke terrain at Wellington to a great depth; but it must be assumed also that there has

been low-latitude immunity from interference by the cryergic processes or that for some other reason running water has remained dominant throughout such alternation of morphogenetic systems as has occurred.

Throughout the Wellington district reliefs of possibly preglacial date and more certainly spur-and-ravine patterns eroded in intercryergic intervals have been eliminated more or less completely by burial under sheets and coulées of head at times when cryergy was active. Many hillslope ravines and minor valleys have been thus completely infilled so that they have vanished. A vast number indeed must still be concealed, as is inferred from the fact that " fossil gullies " (Plate V, fig. 3) are exposed in section wherever cuts are made for roads along hillsides. Thus it is obvious that at all places where the average hillside slopes have not been too steep for fossilization by accumulation of head, such sharp relief as now exists (Plate V, figs. 1, 2) is there not because any very large part of it has survived despite plastering with soliflual debris in cryergic ages but rather as a result of redissection in later times after earlier relief had been thus obliterated.

The foregoing conclusions as to the fast tempo of erosion by hillslope dissection are confirmed by the observations of Dr. H. W. Wellman (personal communication) in the Wairau Valley: (1) Untilled soils on dissected hillslopes are less depleted of plant nutriment than those on even low terraces; (2) Whereas on a terrace 60 feet above river level and probably ten to twenty thousand years old strike-slip displacement of 200 feet on the Wairau Fault is indicated by geomorphic evidence, on a dissected hillslope crossed by the surface trace of the fault offsetting of relief features can be detected only to the extent of about 40 feet, which seems to show that the dissection pattern on the hillslope is changing very rapidly, the present form having been in existence for only some two to four thousand years. This dissected terrain is similar to that at Wellington.

In Europe, where the land surface has presumably escaped dissection in the intercryergic intervals, the broad slopes have become progressively smoother until eventually (in the latest cryergic age) a smooth sheet of head has mantled it instead of one with a corrugated surface made up of confluent coulées. At Wellington, however, where dissection has taken place in the intercryergic intervals, and rather for this reason than because the intensity of cryergy has been less, the smoothness of slopes can at no stage have been as perfect. On imperfectly subdued slopes consequent runnels must develop in all the furrows as soon as streams of water can flow, so that renewed dissection will develop in a pattern quite like that which existed in an earlier age, though inversion of the relief will be common (Cotton and Te Punga, 1955*b*, fig. 1). Thus, while the fossil gullies of the Wellington hills owe their origin to the presence of fine-textured relief on the

major hillslopes that became mantled with head, their eventual exposure is due to a renewed dissection by streams that are largely consequent on the slopes cryergically plastered and to some extent smoothed. It is because such fine-textured dissection and redissection have rarely if ever occurred in Europe that comparable fossilization of relief is rare there, if it occurs at all, as pointed out by Tricart (*supra*).

Mechanism of Fine-textured Sculpture

The mechanism of primary fine-textured dissection and redissection after a cryergic subjugation of the relief cannot be very different. The small streams of water that incise the first formed ravines must be consequent on the general hillside slopes. On a slope corrugated by confluent coulées of head consequent streams will flow in any furrows there are between these on the constructional surface. The dendritic pattern of drainage that has been developed by the time dissection is mature indicates that many insequent branches develop and extend headward from the primary consequents.

It has been claimed that small streams will not corrade vertically as large rivers do, being " incompetent " to carry or roll along large stones, which will therefore be selectively abandoned so as to remain as a pavement in the stream bed, armouring it against corrasion (Tricart, 1956*b*, p. 149). Obviously, however, such an inhibition of vertical corrasion has not operated to prevent dissection of the Wellington landscape. The sharp V section of every one of the innumerable ravines that gash the landscape indicates the fact of its rapid vertical incision, while the complete coverage—the " maturity " of the dissection—demonstrates that headward extension has taken place in most of the ravines, both insequent and primarily consequent.

As admitted by Cotton (in press), the rate of incision by such small streams might be extremely slow in fresh resistant rocks, but where the rocks are no longer resistant owing to deep weathering that has preceded incision of young ravines dissection can be very rapid. The depth and the late date (postcryergic and intercryergic, as indicated above) of the sculpture at Wellington and the feral form of the higher interfluves (despite weakening of rocks by weathering) testify to the efficiency of the running-water morphogenetic system in a humid-temperature climate. In New Zealand, and elsewhere also, such dissection has taken place under forest. This has been under temperate-zone forest, however ; under equatorial rain forest, on the other hand, Tricart and Cailleux (1955, p. 114) point out that chemical (and biological) weathering may proceed so far that no rock fragments remain unrotted. Thus no gravel can be formed, and without gravel as tools of corrasion small streams have no power to corrade vertically. It is only in somewhat cooler climates, therefore, generally in extra-

tropical zones at the present day, where chemical weathering though active enough to provide an easily eroded regolith does not inhibit the formation of gravel, that the process of dissection of the landscape to a fine texture can operate. It is uncertain whether such consequent-insequent dissection is more vigorous under forest than on the slopes of the salient and residual mountain forms of semi-arid regions ; but it must not be lost sight of that many regions now under a semi-arid morphogenetic system were forested in some ages of the Pleistocene.

It has been claimed that any forest protects the land on which it grows from dissection—dissection as distinguished from landsliding and the streaming of a wet regolith in a kind of equatorial solifluxion which it is known may be active. Such claims, however, are based generally on a confusion in their authors' minds of fine-textured dissection by corrading streams as herein described with badland erosion, which is observed to occur after deforestation and from which standing forest has obviously afforded protection, mainly by minimising raindrop-impact effects. In contrast with rainwash such as scours the surface into badlands, however, the streams that incise and carve the landscape in the process of normal fine-textured dissection tap ground water, roll gravel, and corrade quite independently of raindrop impact and superficial scour.

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EXPLANATION OF PLATE V

FIG. 1.—Mature dissection of a strongly warped surface, formerly a peneplain, at Wellington, New Zealand. At right a domical uplift dissected to pyramidal form (Mount Hawkins). View looking SSW. Photo, D. W. McKenzie.

FIG. 2.—Tectonic forms and fine-textured mature dissection of major slopes at Wellington, New Zealand. View looking south over Porirua. Photo, N.Z. Ministry of Works.

FIG. 3.—A " fossil gully " exposed in section, Wellington–Porirua Motorway, New Zealand. As shown by radiocarbon dating of material in the bottom of an adjacent fossil gully, the infilling with geliflual debris took place less than 21,000 years ago. Photo, M. D. King.

FIG. 1.

FIG. 2.

FIG. 3.

EROSION FORMS NEAR WELLINGTON, N.Z.

The Pre-Tertiary Geology of the Western Approaches to the English Channel

By A. A. Day

Abstract

Geological information obtained by core sampling of the solid rock underlying the loose deposits on the floor of the Western Approaches is set forth. Discoveries of interest include those of Chalk in the area north-west of Cornwall and of nepheline phonolite from Epson Shoal, 6 miles east of the Wolf Rock. Other samples obtained very probably represent the Devonian, New Red Sandstone and Chalk, and on the basis of the identifications a geological map has been prepared showing the approximate distribution of these formations in the area.

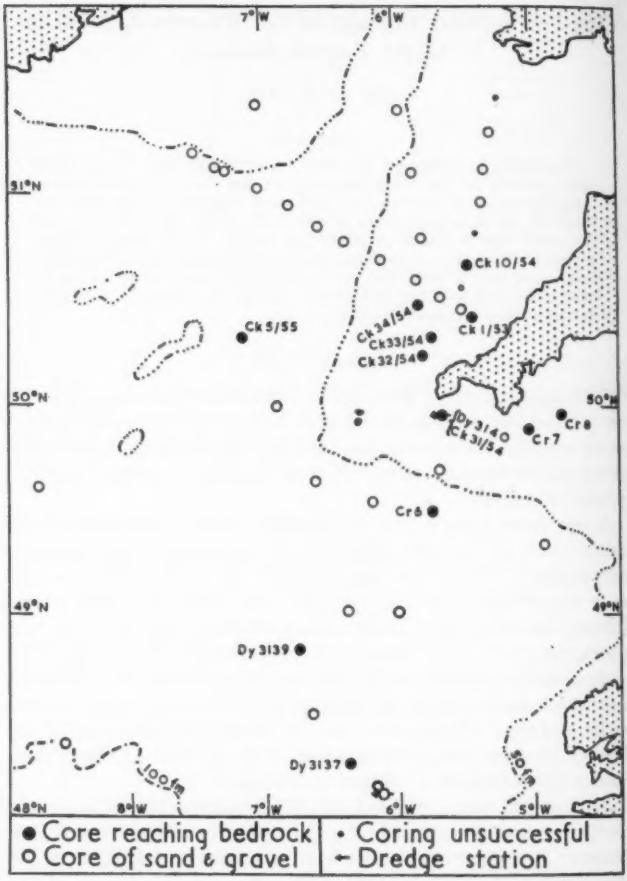
Introduction

THE solid geology of the English Channel has been investigated in considerable detail by King, Hill, and Dangeard. A synthesis of all the available information has been given by King (1954) in a paper that is an outstanding contribution to the much neglected study of submarine geology.

A line from Land's End to Ouessant marks approximately the western limit of the area within which the bedrock has previously been studied; west of this line, in the Western Approaches, as well as in the northern part of the Celtic Sea (between Cornwall and Ireland), the loose late Tertiary and Quaternary deposits have such a thickness that it is not normally possible to penetrate through them to the underlying bedrock with the corers at present in use. However, a few cores have brought up samples of older strata in the western waters, and the rock exposed in several submarine banks lying off the Cornish coast has also been sampled. The positions at which these samples were obtained are shown in Text-fig. 1.

The samples were collected by H.M. Survey Ships *Cook* and *Challenger*, and by Dr. M. N. Hill and the author on R.R.S. *Discovery II*, the respective ships being indicated in Text-fig. 1 by the abbreviations Ck, Cr, and Dy. The corers employed were of the free-fall type described by Hill and King (1953). The gross weight of the corers used on the three ships was dependent on the capacity of the winches and the wire available on each. On *Discovery II* it was possible to use a corer weighing about 900 lb. The penetrating power of this corer in areas of the sea floor covered by loose sediment was accordingly greater than that of the lighter corers normally used by the survey ships.

The positions of the stations at which either the core obtained did not penetrate to the solid rock or no core was obtained at all, are



TEXT-FIG. 1.—Map showing position of the eleven coring stations.

also shown in Text-fig. 1. In the present area the failure of a coring station may with reasonable certainty be ascribed to the incoherent nature of the surface deposits on the sea floor for, if the corer does not penetrate to a more coherent layer, the loose surface material may escape through the core-catcher at the bottom of the core-barrel and be lost. It is possible, however, that a failure might indicate hard, bare rock, or a submerged wreck, especially if the hardened steel cutting edge of the corer is found to have been chipped.

The Core Samples

The nature and probable geological age of the pre-Tertiary samples collected at eleven coring stations are set out in the accompanying table. The station numbers correspond to those shown in Text-fig. 1.

Notes on Stations Dy 3137 and 3139.—The chalky material obtained at station Dy 3137 was very soft when collected, but dried extremely hard. Since the material was so soft in its waterlogged condition it is unlikely that it could have resisted the abrasion which it would have suffered if it had been transported by bottom currents from, say, the English Channel. A more plausible assumption is that the sample represents chalk in place.

Dr. J. M. Hancock, of King's College, London, very kindly examined the specimen. He expressed the opinion that lithologically it closely resembled the chalk of the type Maestrichtian in the Netherlands. Further, Dr. Hancock stated that polyzoa present in the sample appeared to belong to the family Membraniporidae (Cretaceous to Recent), but they were not sufficiently well preserved for generic identification.

At station Dy 3139 a small amount of soft white mud was found adhering to the steel cutter of the corer and diffused in the lowest 1 cm. of the loose gravel and shells held in the core-barrel. This white mud is considered to represent a soft chalky layer underlying the gravel, and not a transported chalk pebble, for the reasons outlined above in the discussion of specimen 3137.

A Cretaceous age is ascribed to the samples from both localities, but the evidence is not sufficient entirely to preclude the possibility of their belonging to a marine Tertiary deposit, especially in the case of specimen 3139.

Discussion of the Coring Results

General.—In attempts to determine the solid geology beneath the mantle of loose deposits by coring, there is clearly inherent the danger that the corer may strike a large cobble or boulder and the sample then mistakenly be considered to have been derived from solid bedrock. Large boulders of the harder rock-types are known to occur on the bottom in parts of the waters surrounding south-west England, and dredge samples from the area commonly contain a number of pebbles which are larger than the diameter of the corer jaw used (2 inches). Arguments have been advanced in the previous section in support of the assumption that the samples were truly representative of the solid bedrock. The results, on this assumption, are not inconsistent with

TABLE 1.—DESCRIPTION OF PRE-TERTIARY ROCK SAMPLES FROM THE WESTERN APPROACHES

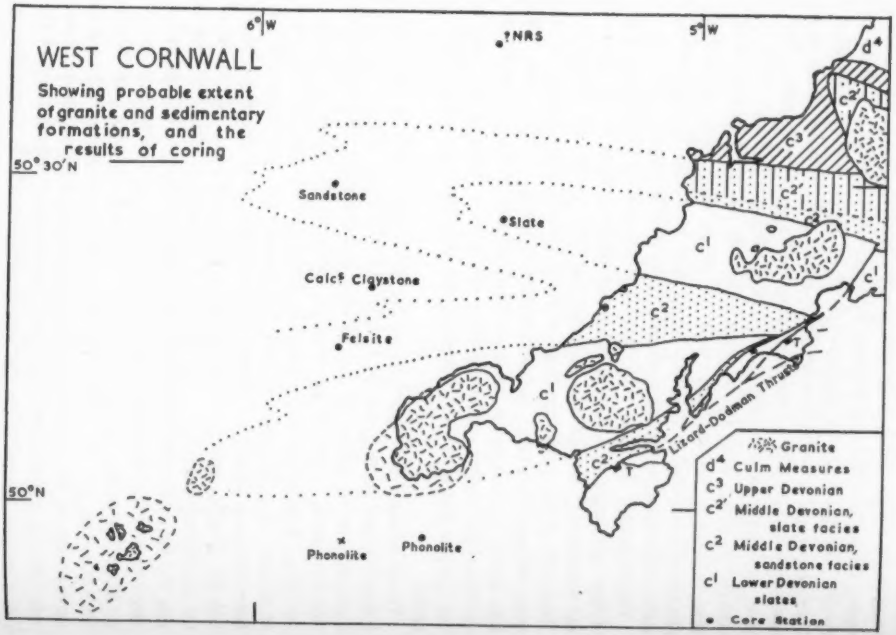
Station No.	*Latitude N.*	*Longitude W.*	*Depth. fm.*	*Length of Core and Position of Sample.*	*Description of Sample and Probable Geological Age.*
Cr 6 . .	49° 30′.0	5° 47′.3	51	15 cm.; 9–15 cm.	White chalk with flints; Cretaceous.
Cr 7 . .	49° 52′	5° 01′	46	20 cm.; 13–20 cm.	Chocolate-brown micaceous mudstone; N.R.S.
Cr 8 . .	49° 58′	4° 47′	42	19 cm.; 13–19 cm.	Red mudstone with small rounded pebbles of limestone and quartzite; N.R.S.
Dy 3137 .	48° 17′	6° 22′	76	32 cm.; 29–32 cm.	White chalk; Cretaceous. See notes in text.
Dy 3139 .	48° 50′	6° 45′	74	9 cm.; very small sample.	White mud; probably Cretaceous. See notes in text.
Ck 1/53 .	50° 26′.4	5° 27′.0	21	21 cm.; 14–21 cm.	Greenish-grey, well cleaved slate; probably Devonian.
Ck 10/54 .	51° 41′.7	5° 28′.1	40	18 cm.; 16–18 cm.	Stiff clay with strong red colouration in lower part. The colour is very suggestive of that of typical shales of the New Red Sandstone, and is rare among the Devonian on the neighbouring land. Probable age: New Red Sandstone.
Ck 32/54 .	50° 14′.4 Bank, 8 miles N.W. of Cape Cornwall.	5° 49′.3	20	Rock fragments	Freshly-broken fragments of acid felsite; greenish, with one direction of cleavage; Palaeozoic.
Ck 33/54 .	50° 19′.6 The Bann Shoal.	5° 45′.2	19	Rock fragments	Freshly-broken fragments of slightly thermally-metamorphozed calcareous claystone; probably Devonian.
Ck 34/54 .	50° 29′.4	5° 50	41	18 cm.; 15–18 cm.	Brown, cleaved micaceous sandstone; probably Devonian.
Ck 5/55 .	50° 20′	7° 09′	60	24 cm.; 10–24 cm.	Soft white chalk; Cretaceous.

the distribution of rock-types that may reasonably be expected in the area.

In an early paper on the English Channel, Austen (1850, pp. 84 and 86) stated that rock was exposed in the Jones Bank (49° 50′ N., 8° W.) and on the edge of the continental shelf, without specifying the type of rock present in either case. These observations by Austen are not corroborated by the detailed bottom descriptions given against the soundings on Close's "Southern Grounds" chart or on the Admiralty charts. Also, an echo-sounding profile taken across the Jones Bank by the author in 1955 shows that the bank has gently-sloping flanks very similar to those of other sand-and-gravel banks in the area; no steep slopes suggestive of rocky exposures were observed. Consequently, it is doubtful whether Austen's conclusions, which were based on observations by the sounding lead alone, are correct; they have, therefore, been neglected in the discussion below. That no outcrops of older beds are at present known to occur in any of the large submarine banks lying well off shore does not, of course, preclude the possibility that the banks may be ridges in the bedrock surface that are covered by a mantle of loose sediment.

The Palaeozoic Samples.—In Text-fig. 2 are shown the solid geology of West Cornwall and the probable submarine extent of the granite and sedimentary formations, together with the results of the coring. The extent of the granite masses of the Isles of Scilly and the Seven Stones is very uncertain; the boundaries shown by broken lines have been drawn to include all the associated submarine knolls shown on the Admiralty charts. The probable southward limit of the Land's End Granite has also been based on the distribution of the offshore knolls.

The generalized solid geology of the mainland shown in Text-fig. 2 is largely based on the map by Hendriks (1937, Plate 22) in which she illustrated the structure of the Devonian formations by postulating their submarine extent in the manner indicated by the dotted lines in the figure. It may be seen in Text-fig. 2 that the sample of slate found at station Ck 1/53 agrees with the postulated occurrence of the Dartmouth and Meadfoot Slates, and the sandstone at station Ck 34/54 with the possible extent of the Staddon Grit and the sandy facies of the Gramscatho Group. An important calcareous slate bed occurs near the base of the Upper Devonian, and the sample of calcareous claystone (Ck 33/54) agrees very well with the postulated position of the Upper Devonian strata. The felsite of the bank near Cape Cornwall, however, is not a rock-type that would be expected to occur in the exposure of the Gramscatho Group that Miss Hendriks suggested may occur in the vicinity. The most reasonable interpretation of the felsite sample seems to be that the bank from which it was derived



TEXT-FIG. 2.—Geology of W. Cornwall and surrounding sea floor (based on E. M. L. Hendriks, 1937).

is part of an elvan possibly associated with the Land's End Granite. The bottom samples at present available are, therefore, not inconsistent with the distribution of the Devonian formations postulated by Miss Hendriks.

New Red Sandstone Samples.—Stations Cr 7 and 8, at which samples of undoubted New Red Sandstone were obtained, fall within the area of outcrop shown by King in his geological map of the Channel (1954, Plate IV), and merely serve to confirm that map. On the other hand, the probable presence of the New Red Sandstone at station Ck 10/54 is an isolated record in an area in which the nearest exposure of the system occurs on the southern shore of Barnstaple Bay, approximately 55 miles distant. This new record suggests that the New Red Sandstone may occur under part, at least, of the northern Celtic Sea, in which case the palaeogeographic reconstructions of Wills (1951) and others for the Permian and Triassic periods will require slight revision. The maps should then show a basin lying south and west of South Wales, perhaps connected to the basin in which were deposited the New Red rocks lying beneath St. George's Channel (Hill, 1955).

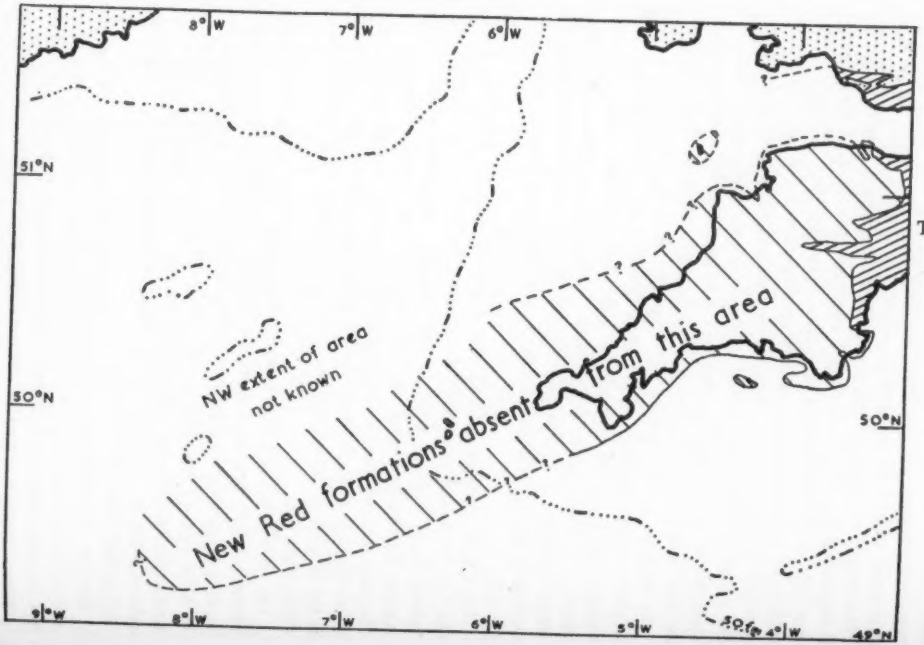
An analysis of the results of seismic prospecting at a number of stations in the Western Approaches (Day *et al.*, 1956, pp. 36, 37) suggests the probable absence of New Red deposits from an area lying to the south-west of Cornwall. Taking the seismic and coring results together, it is possible to delineate the approximate limits of outcrop of the New Red Sandstone on the Palaeozoic floor in this region (Text-fig. 3).

The Chalk Samples.—Chalk was found at stations Dy 3137, Cr 6, and Ck 5/55 and probably at station Dy 3139. The samples are derived from a wide area of the Western Approaches and much increase the known area of occurrence of the Chalk (Text-fig. 4). This map summarizes all the available information on the distribution of pre-Tertiary formations in the Western Approaches; the boundaries in the English Channel, east of 5° W., have been adopted from King (1954, Plate IV).

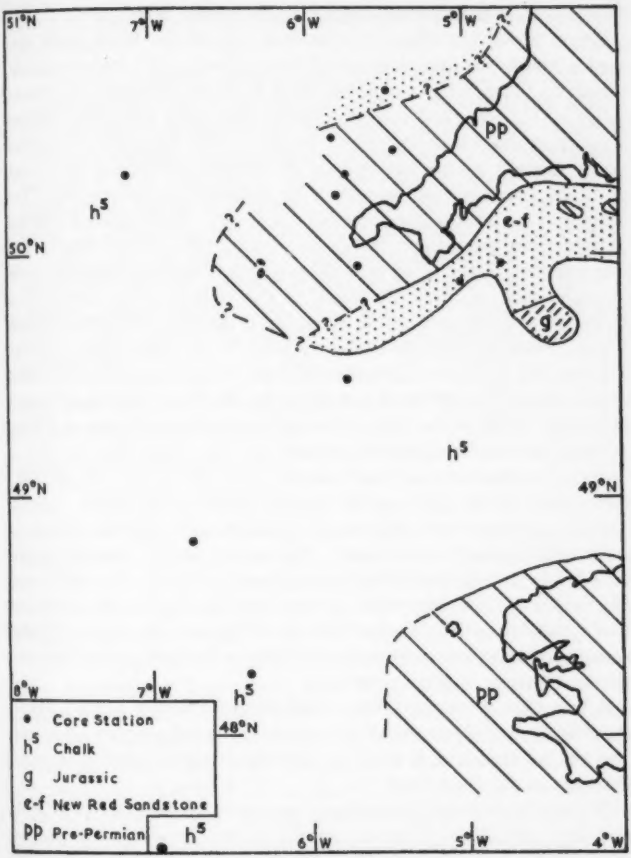
The record of chalk north-west of the Scilly Isles is the first for the northern Celtic Sea, and suggests that the formation may be present under a large part of this area. The chalk pebbles and flints found by Cole and Crook (1910, pp. 21–6) in dredge samples from the continental slope west of County Kerry may perhaps have been derived from an extension of the Chalk formation from the Celtic Sea to the shelf west of Ireland.

Text-fig. 4 shows the Chalk overlapping all the older formations, and indicates that the submarine continuations of the outcrops of the pre-Permian rocks of Cornwall and Brittany are limited in extent.

There remains a part of the sea floor, at the outer margin of the



TEXT-FIG. 3.—Map showing approximate limits of the New Red Sandstone.



TEXT-FIG. 4.—Submarine outcrops in the neighbourhood of Cornwall and Brittany.

continental shelf, concerning the solid geology of which we have no information. Attempts to penetrate with the corer to older material than the Quaternary sand have hitherto been unsuccessful, but the thinness of the sand and gravel mantle in the area adjacent to the mouth of the English Channel raises the hope that this mantle may also be thin near the shelf margin and that the older strata may yet be accessible to the corer.

Coring and Dredging on Epson Shoal

Epson Shoal lies about 6 miles due east of the Wolf Rock and approximately the same distance to the south of Land's End peninsula. It rises to 14 fm. from a regional depth of 36 fm. (implying an overall height of 132 feet), and according to the Admiralty charts the bottom is rocky at the shallowest point. In July, 1954, *Discovery II* steamed slowly along a zig-zag course over the plotted position of the shoal, using the Decca navigation system for accurate position-finding. The shallowest echo-sounding obtained was 20 fm. Wind made it difficult to manoeuvre the ship properly and we may have passed first to one side and then to the other of a sharp pinnacle without detecting it on the echo-sounder. On the basis of the rough survey made, it is estimated that the base of the shoal has a diameter of approximately ¼ mile, while the portion shallower than 20 fm. must have a width less than 500 feet. It is also apparent from the echo-sounding traverses that a sharp decrease in the slope of the flanks of the shoal occurs at a depth of 28–30 fm. This feature suggests the existence of a heap of loose sand and gravel piled against the base of the shoal.

A station, Dy 3140, was worked in a position 49° 57′.8 N., 5° 38′.9 W., and a depth of 22–7 fm., on the northern flanks of the shoal. Dredge and core samples were obtained, but these were composed entirely of loose gravel, shells, and sand. The gravel in the dredge sample included large pebbles of flint, quartz, granite, slate, and diorite with one pebble of phonolite. None of these pebbles showed any indication that it had been freshly broken from an outcrop on the shoal. Neither were any rock-types represented among the pebbles which were not known to occur on the nearby land.

A thin slice of the phonolite pebble showed the rock to be a microporphyritic nepheline phonolite with a very small amount of leucite. The pebble, therefore, is of a type slightly different from that forming the mass of the Wolf Rock.

H.M.S. *Cook* carried out coring at a position 49° 57′.2 N., 5° 38′.8 W. in a depth of 22 fm. This station, Ck 31/54, lies about 0·6 nautical mile slightly east of south from the *Discovery* station. The sample obtained, using an 800 lb. corer, consisted of a few polyzoan fragments and several angular pieces of phonolite with freshly broken surfaces. The shape and condition of the rock fragments suggests that they were derived from the solid rock of the shoal at this point. The rock is megascopically very similar to the phonolite pebble found at station Dy 3140. In thin section it is seen to be a microporphyritic nepheline phonolite.

Thus, the evidence suggests that the Epson Shoal is a companion phonolitic mass to the Wolf Rock, and forms the second known occurrence of this rock-type in the south of England. Owing to the

interest of the discovery a petrographic description of the specimen is appended:—

> The rock is slightly altered, but appears originally to have been holocrystalline, with a sub-ophitic texture. Sanidine forms tabular phenocrysts commonly altered to sericite along curved transverse fractures, and also is present in the groundmass in conjunction with nepheline and aegerine. Nepheline occurs rarely as microphenocrysts, but forms approximately 60 per cent of the groundmass, in which it is present in clear hexagonal or square crystals, 0·5–0·1 mm. in diameter, occasionally enclosing a few small, irregularly oriented microlites. The aegerine forms sub-tabular to acicular green microlites, commonly between 0·05 and 0·1 mm. in length, uniformly distributed and randomly oriented throughout the groundmass.

The exact nature and structural relations of the phonolitic masses forming Wolf Rock and Epson Shoal are not clear. They may represent the cores of small pipe-like intrusions from which the surrounding Palaeozoic country rock has been removed by erosion during a period of lowered sea-level, or they may be the remnants of flows extruded in the relatively recent past (for example, the late Tertiary). A measurement of the gravity anomaly on the Wolf Rock lighthouse may possibly indicate the size of the underlying phonolitic body, for the density of the phonolite (2·55 gm./cc.) is appreciably less than that of either granite or the Palaeozoic country rock.

The specimen Ck 31/54 differs notably from the Wolf Rock specimens described by Teall (1888, p. 367) in its lack of nosean. This difference is relatively unimportant, however, for it could have arisen from slight variations in the composition of a single magma source.

From their close proximity it is probable that the two rock masses are genetically related. They lie slightly to the north of the westward extension of the Lizard thrust, but there is no strong reason for supposing that they are genetically related to the thrusting. No New Red Sandstone is at present known to outcrop in contact with the phonolite masses, so that the time of their formation remains indeterminate.

Acknowledgments

I am grateful to the Hydrographer of the Navy for permitting me to examine the cores collected by H.M. survey ships; the officers of these ships have carried out the difficult operations of coring with great skill. To Dr. M. N. Hill, Dr. J. C. Swallow, Professor W. B. R. King, and Dr. S. O. Agrell I am indebted for advice and assistance.

The work was carried out in the Department of Geodesy and Geophysics, Cambridge.

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Age of Zircon from the Kunait Syenite Complex, South-West Greenland

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Abstract

The age of zircon from the Kunait pluton in south-west Greenland has been measured by the lead-alpha method. The experimental data indicate a Cambrian age and the bearing of this result on the general geology of the area is briefly discussed.

Introduction

THE greater part of the ice-free coastal belt of south-west Greenland consists of an ancient, highly-folded basement complex composed mainly of high-grade gneisses. Rocks of a somewhat lower metamorphic grade, however, occur in restricted areas. These are known as the Arsuk series and their original sedimentary nature is clearly recognizable. Granitic rocks, e.g. the Julianehaab granites, also form part of the basement complex.

The basement complex is overlain unconformably in the Julianehaab region by the Gardar formation which comprises a series of unaltered, scarcely folded sandstones, succeeded by volcanic horizons. The sandstone and lava sequence is devoid of fossils, apart from some non-diagnostic sponge remains and has been variously considered to be Eo-Cambrian (comparable with the Torridonian of north-west Scotland), Devonian, or even Permo-Triassic. Ussing (1912) regarded the Gardar formation as Devonian. On this assumption he referred the spectacular nepheline-syenite and soda-granite complexes of Ilimausak and Igaliko, which are intruded into the Gardar formation, to a late phase of the Devonian.

From a comparative study of the available geological evidence, Wegmann (1938), on the other hand, considers the most probable age of the Gardar formation to be young Pre-Cambrian or Palaeozoic. This is essentially in agreement with the result presented in this paper, with the assumptions stated below.

The alkaline masses Ilimausak and Igaliko belong to a group of "Newer Intrusives". The Kunait Complex is one of the smaller Newer Intrusives and occurs near Ivigtut (Wegmann, 1938; Callisen, 1943; Bondam, 1956). The Kunait syenites cut the basement gneisses approximately 150 km. west of the Ilimausak–Igaliko region where the nearest Gardar rocks outcrop. It is probable that the Kunait syenites are contemporaneous with those of Ilimausak and Igaliko and an age determination on Kunait should, therefore, give a minimum age for the Gardar sedimentary series.

The present paper is concerned with the age of zircons from a pegmatite vein within the Kunait pluton.

Experimental Methods and Results

An age-determination on the Kunait zircon was carried out by the "lead-alpha" method proposed by Larsen *et al.* (1952), which has been successfully applied to zircons by a number of workers in recent years (Larsen *et al.*, 1953; Gottfried *et al.*, 1956; Webber *et al.*, 1956; Roques, 1956*a*). The assumptions made in the method are that all the lead in the zircon is radiogenic and that no loss or gain of radioactive elements or lead has occurred during the geological history of the mineral. In many cases this method gives ages which agree to within 10 per cent of the values obtained by a complete isotopic analysis (Gottfried *et al.*, 1956).

The zircons were easily separated by hand-picking of the coarsely crushed pegmatite. The dark brown crystals were crushed to pass 20-mesh and any compound grains of zircon and feldspar removed with the aid of a binocular microscope. The lead content was determined spectrographically by the method of Waring and Worthing (1953), except that indium was used as internal standard. The total alpha-activity of the sample was measured by a simple thick-source technique, using a zinc sulphide-screen scintillation counter. All instruments were calibrated with several standard zircon samples of known uranium, thorium, and lead content, supplied by Dr. H. D. Holland, of Princeton University, U.S.A., at present a National Foundation Research Fellow at the Department of Geology and Mineralogy, Oxford University. An X-ray photograph of the powdered material disclosed no lines other than those of zircon.

The results of the measurements were as follows :—

Lead (micrograms per gram of zircon) : 144, 128, 130, 125, 120, 146, 124, 130. Mean : 131.

Alpha counts (per milligram zircon per hour) : 619 ± 25 (mean of numerous measurements).

The approximate formula for calculating the age is :—

$$t = \frac{c\,Pb}{\alpha} \quad . \quad . \quad . \quad . \quad (1)$$

where t is the age in millions of years, Pb is the lead content in micrograms per gram, α is the alpha-activity per milligram per hour, and c is a constant depending on the ratio of uranium to thorium in the sample. Roques (1956*b*) has conveniently given a table of figures for the preparation of a series of graphs by means of which it is possible to read off the apparent age for different uranium/thorium ratios as a function of the lead content and the total alpha-activity.

The uranium/thorium ratio of the Kunait zircon has not yet been determined. Table 1 gives the ages for various values of this ratio, using the above experimental data.

F is defined as :—

$$\frac{\text{Amount of uranium present}}{\text{Amount of uranium required to yield the same number of alpha-particles as the sample}}$$

Larsen, Keevil, and Harrison (1952) give an average value of $F = 0{\cdot}71$ for zircons. Using this value for the Kunait zircon, its age comes out as 497 million years.

TABLE 1 *

F	0·00 (all thorium)	0·20	0·40	0·60	0·71 (average zircon)	0·80	1·00 (all thorium)
Age in m.y.	414	437	456	481	497	508	538

* Determined graphically.

The Kunait syenite thus appears to be of Cambrian age, as judged by the time scale of Holmes (1946). On the assumption that the Ilimausak–Igaliko complexes were intruded contemporaneously, this also gives a minimum age for the Gardar formation. It is intended to undertake the isolation and isotopic analysis of the lead in Kunait zircon in the near future.

This work has been undertaken as part of a programme of geological age and isotope studies in the Department of Geology and Mineralogy, Oxford University. Grateful acknowledgment is made to Grønlands Geologiske Undersøgelse, Copenhagen, under whose auspices the material was collected in summer, 1956.

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On Post-Glacial Pelicans in Britain

By C. L. Forbes, K. A. Joysey, and R. G. West

Abstract

A bone found in the peat near King's Lynn, Norfolk, has been identified as belonging to a pelican. This is the fifth fossil record from East Anglia but it is the first to be accurately dated, and it can be correlated with Godwin's pollen zone VII–VIII (Iron Age). Other fossil pelican bones from East Anglia and Somerset are identified as those of the Dalmatian Pelican rather than the White Pelican which now visits parts of north-west Europe.

Introduction

THE pelican bone described in this paper was collected by Mr. W. E. Doran, Chief Engineer of the Great Ouse River Board, from the new flood relief channel near King's Lynn, Norfolk. It was found about half a mile north of Saddle Bow village, National Grid Reference TF 608166. The Great Ouse River Board have kindly presented it to the Sedgwick Museum, Cambridge.

We are indebted to Mr. J. D. Macdonald and Mrs. P. A. Cottam for their help in the British Museum (Natural History), to Dr. M. Bird and Miss M. T. Prior for arranging the loan of specimens from the Glastonbury Lake Village Museum, to Miss A. Pringle and Mr. R. D. Norman for their assistance with the typescript and illustration respectively, and to Mr. E. W. Dawson for fruitful discussion and criticism of the manuscript.

Description

The present specimen (Sedgwick Museum D.5762) consists of three fused thoracic vertebrae, which have been broken from the anterior end of the synsacrum. The centra and neural arches are well preserved but the transverse processes are incomplete.

For the purposes of comparison eight complete skeletons of pelicans have been examined and in all of these the first and second thoracic vertebrae are separate, while the subsequent ones are fused. The constant position of the limit of fusion indicates that the most anterior vertebra in the present specimen is the third thoracic. The structure of the anterior region of the synsacrum is extremely variable due to individual differences in the degree of ossification. It is noteworthy that the spinal nerve foramina of the present specimen are larger than those in any of the other specimens which have been examined, but this minor difference does not warrant the foundation of a new species.

At present there are two species of pelican living in Europe, the Dalmatian Pelican (*Pelecanus crispus* Bruch) and the White Pelican

(*Pelecanus onocrotalus* Linné). The anterior region of the synsacrum does not provide any characters on which it is possible to distinguish between these two species, and so the present specimen must remain *Pelecanus* of uncertain species.

Stratigraphy, Correlation, and Palaeoecology

The section of Fenland post-glacial deposits exposed where the pelican bone was found is as follows :—

4. Grey silt, weathering brown above, penetrated from the surface by *Phragmites* rhizomes. 0 to 3 feet.
3. Chocolate-brown fen peat with wood and *Phragmites* rhizomes. 3 ft. to 5 ft. 6 in.
2. Transition to next. Brown clay penetrated by *Phragmites* rhizomes. 5ft. 6 in. to 5 ft. 11 in.
1. Grey-blue fen clay, with *Cardium edule* abundant in the top 6 in.; seen to water level. 5 ft. 11 in. to 17 ft. 11 in.

Table 1 shows the relation of these deposits to the general succession of Fenland deposits, pollen zones, and archaeology described by Godwin (1940). The sequence—fen clay, peat, silt—is repeated in the section described by H. and M. E. Godwin (1933) at Wiggenhall St. Germans, about a mile and a quarter south of the pelican site. Our peat bed No. 3 corresponds to the 2 ft. peat bed " E " at Wiggenhall St. Germans.

Table 1.—Correlation of Deposits (after Godwin, 1940)

Bed.	*Pollen Zone.*	*Archaeology.*	*Approximate Age.*
4. Silt . . .	VIII	Romano-British	0 A.D.–B.C.
3. Peat . . .	VII–VIII	Iron Age	500 B.C.
(2. Transition)			
1. Fen clay . .	VIIc and *d*	Bronze Age	

The pelican bone was stained brown and contained peat in its crevices and hollows, so there is no doubt that it came from the peat bed, although it was found lying on the peat in a slumped part of the section. In order to find the horizon within the peat from which the bone came, a pollen analysis of the peat scrapings from the bone was made, and the results of this analysis, given in Table 2, were then compared with other analyses made from successive levels of the peat bed itself. The pollen analysis of peat from the bone was also compared with the analyses made by H. and M. E. Godwin (1933) through the synchronous 2 ft. peat bed at Wiggenhall St. Germans.

Samples were taken at 6 in. intervals through the peat bed a few yards from the pelican site. The samples from the top 2 feet of the bed contained sparse tree pollen, but abundant fern spores, values of which reached from 600 to 6,000 per cent of the total tree pollen.

The results indicate that the bone came from the lowest 6 inches of the peat bed, where values for the fern spores were found to be considerably lower and tree pollen much more abundant. A similar horizon for the provenance of the bone is suggested by comparison with the pollen diagram from Wiggenhall St. Germans; here the analysis from the pelican bone is best matched in the section of the diagram taken from the lower part of the peat bed.

The general ecological surroundings indicated by the section may now be considered. As described by Godwin (1940), the fen clay was formed under brackish-water conditions, the freshwater 2 ft. peat bed was deposited after a regression of the sea, and the change to the upper silt was caused by a later marine transgression. H. and M. E.

TABLE 2.—POLLEN ANALYSIS OF PEAT SCRAPINGS FROM THE PELICAN BONE
Percentages of Total (150) Tree Pollen.

TREES			
Betula (Birch)	20	*Tilia* (Lime)	1
Pinus (Pine)	1	*Alnus* (Alder)	27
Ulmus (Elm)	6	*Carpinus* (Hornbeam)	1
Quercus (Oak)	36	*Fraxinus* (Ash)	8
NON-TREES			
Corylus (Hazel)	62	*Ranunculus* (Buttercup)	1
Salix (Willow)	3	Umbelliferae (Umbellifers)	1
Gramineae (Grasses)	29	*Myriophyllum verticillatum* (Whorled water-milfoil)	4
Cyperaceae (Sedges)	12	*Nymphaea* (Water-lily)	3
Artemisia (Mugwort)	3	*Sparganium* (Bur-reed)	20
Calluna (Ling)	1	*Typha latifolia* (Great reedmace)	1
Chenopodiaceae (Goosefoot family)	3	Filicales (Ferns)	55
Rubiaceae (Bedstraw family)	1	*Pteridium* (Bracken)	1
Plantago lanceolata (Ribwort)	3		

Godwin (1933) have described the vegetational conditions obtaining during the growth of the peat in this area, as indicated by their pollen diagram from Wiggenhall St. Germans. The basal part of the peat contains indicators of salt-marsh conditions (e.g. the 3 per cent of Chenopodiaceae pollen seen in the analysis from the pelican bone). Such conditions would appear on the regression of the sea at the change from the fen clay to the fen peat. The lower middle part of the peat was formed during alder carr conditions, when brackish-water gave way to freshwater conditions. In the analysis from the bone the presence of shallow open freshwater is suggested by the presence of pollen of *Myriophyllum verticillatum* and *Nymphaea*, and of reed swamp by *Typha latifolia* and *Sparganium*. The remaining upper part of the peat was formed under fen wood conditions. This change from alder carr to fen wood is the normal vegetational succession to be expected. Thus, the pelican bone came from the vegetational stage

during the replacement of the salt-marsh by freshwater fen, and prior to the growth of the fen wood.

Finally, we may note that there is no clear evidence from this site of former climatic conditions or changes which might have affected the distribution of the pelican.

Previous Records of Fossil Pelicans

Milne-Edwards (1868) and Newton (1868*a*) recorded the first pelican bone from Britain. The bone (Sedgwick Museum No. D.5760) was among an old collection of bones from the lower peat of Cambridgeshire. It is a left humerus, which was judged to have been derived from a young bird, presumably bred in this country, rather than from an adult migrant. Newton stated that it was larger than the corresponding bone of the White Pelican, *Pelecanus onocrotalus* Linné, and on the same basis Milne-Edwards suggested that it might belong to a new species.

In 1871, Newton described another pelican humerus from Feltwell Fen, Norfolk. This bone was presented by J. H. Gurney to the University Museum of Zoology, Cambridge (No. 260*a*). Newton noted that it was from an adult bird, and suggested that both this and the previous humerus were comparable in size to that of the Dalmatian Pelican, *Pelecanus crispus* Bruch.

Harmer (1897) recorded the discovery of the fractured, but associated, humerus, radius, and ulna of a pelican from Burnt Fen, near Littleport, Cambridgeshire. These bones, which are also in the University Museum of Zoology, Cambridge (No. 258D), were identified as belonging to *Pelecanus crispus*.

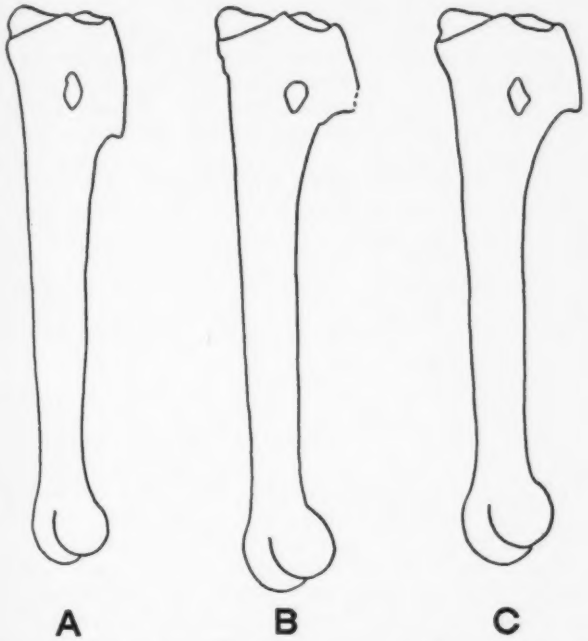
Newton (1901) recognized a right tarso-metatarsus of a pelican among the collection of bird bones in which he made his first discovery and he suggested that they might have been collected from the same deposit. He also identified this bone (Sedgwick Museum No. D.5761) as *Pelecanus crispus*.

In the course of the present study all of the pelican bones previously found in East Anglia have been re-examined and compared with skeletons of *P. crispus* and *P. onocrotalus*. The seven skeletons of *P. onocrotalus* available for study in the British Museum fall into two size groups, five of the skeletons being distinctly smaller than the other two. The sex of only two of the specimens is definitely known and these are both female, and both belong to the smaller size group; this is consistent with the assumption that the differences in size may be due to sexual dimorphism.

Only a single complete skeleton of *P. crispus* is available for study, and its sex is unknown. The bones of the wing are comparable in size with the two larger (male ?) skeletons of *P. onocrotalus*, but the

leg bones are relatively small and comparable in size with those of the smaller (female) skeletons of *P. onocrotalus*.

All the wing bones of the pelicans found in the peat of East Anglia are comparable in size to the skeleton of *P. crispus* and the two



TEXT-FIG. 1.—Right tarso-metatarsi of pelicans, viewed from the left side. ($\times \frac{3}{4}$.)

A. *Pelecanus onocrotalus* (British Museum, 1903.3.6.1).
B. Fossil pelican from East Anglian peat (Sedgwick Museum, D.5761).
C. *Pelecanus crispus* (British Museum, 1896.2.7.1.)

male (?) skeletons of *P. onocrotalus*. These fossil wing bones were originally referred to *P. crispus* on the basis of their large size, but it is evident that some specimens of *P. onocrotalus* attain an equivalent size, and one of them (British Museum No. 1903.3.6.2) is larger than any of the East Anglian fossils.

It has not been possible to find any characters of the humerus, radius, or ulna which distinguish *P. crispus* from *P. onocrotalus*, and

so all the East Anglian fossil wing bones should strictly be referred to as *Pelecanus* of uncertain species.

The remaining pelican bone previously found in the East Anglian peat is the tarso-metatarsus recorded by Newton (1901). The tarso-metatarsus of the single skeleton of *P. crispus* in the British Museum shows differences from those of *P. onocrotalus*. In *P. crispus* the hypotarsus projects relatively further and its posterior face is relatively smaller than in *P. onocrotalus* (Text-fig. 1). As there is only one skeleton of *P. crispus*, it seemed possible that these differences were due to individual variation. Fortunately, when a bird skin is prepared as a study specimen the leg bones are usually left in position. A single specimen (British Museum No 1895.2.10.112) was chosen at random from the large collection of skins of *P. crispus*, and the tarso-metatarsus was carefully exposed. It was found that the shape of the hypotarsus matched that of the single skeleton of *P. crispus*, so confirming that this character could be used to distinguish between the two European species of pelican. As a further check, the tarso-metatarsus in the skin of an immature specimen (British Museum No. 1908.2.29.1) of *P. onocrotalus* was similarly exposed, and it was found to conform to the type expected in this species.

The tarso-metatarsus from the peat of East Anglia is of the type found in *P. crispus*, so confirming the conclusion reached by Newton.

Elsewhere in Britain, a single femur of a pelican has been found in the peat of Hull, Yorkshire (E. T. Newton, 1928), and the remains of *P. crispus* have been recorded from the Lake Village at Glastonbury, Somerset, which was the site of an Iron Age settlement shortly before the Roman occupation. Andrews (1899, 1917) described numerous bones, including nine tarso-metatarsi, so representing at least five individuals, some of which were young birds. It has been possible to re-examine eight of the Glastonbury tarso-metatarsi. Five of these bones have the hypotarsus preserved and all of them are of the *P. crispus* type.

In an extensive study of references to birds in early literature, Gurney (1921) found no evidence that the pelican was native to Britain either during or subsequent to Roman times. This indication is supported by the absence of the pelican among bird bones which have been recovered from sites of Roman occupation in Britain.

Records of Living Pelicans in Britain

The most recent check-list of the birds of Great Britain and Ireland (1952) does not include any species of pelican, but there are several reports of pelicans having been shot or sighted in England during recent times. Sir Thomas Browne recorded that a pelican was shot at Horsey Fen, Norfolk, in 1663, but he suggested that it might have

been one of the King's pelicans which had escaped from St. James Park at about the same time (Newton, 1868*b*). Nearly two hundred years later, Tristam (1856) recorded that a dead pelican had been found on the coast of County Durham, but gave no indication of the species.

Gurney (1907) reported a pelican at Breydon, Norfolk, which later spent several months near Whitstable, Kent, where it was identified as a White Pelican, *P. onocrotalus* (Saunders, 1906). Another pelican was sighted at Breydon nine years later (Patterson, 1916) but this time the species was not recorded.

According to Dresser (1881) the ranges of *P. onocrotalus* and *P. crispus* overlap to a large extent. Both species live in South and Southeastern Europe and range into North Africa and eastwards into Asia. *P. onocrotalus* occurs rarely in Central and Northern Europe although it has been recorded in North Germany, Poland, Denmark, Sweden, and Finland. In contrast, *P. crispus* has not been observed either in Scandinavia or France.

On the basis of this present distribution, supported by the evidence of the only specimen which has been identified, it seems likely that the stray pelicans which have been recorded in Britain during the last few hundred years have been *P. onocrotalus* and hence a different species from that which occurs in the Iron Age peat. Finally, it is worthy of note that Winge (1901) has reported the occurrence of fossil *P. crispus* in Denmark, which is also outside the present range of the species.

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High-level Gravels in and near the Lower Severn Valley

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Abstract

The Woolridge Terrace gravels of the lower Severn are re-examined, and certain high-level gravels west of the Severn valley examined for the first time. It is concluded that the deposits are of two different ages. The older series is probably the outwash of a Severn valley glacier, the younger was laid down by a stream coming from the Silurian hills near Ledbury. A Gipping age is suggested for both.

I. Introduction

In the lower Severn valley Wills distinguished five principal aggradation terraces, of which the highest and oldest was the Woolridge. It was represented by small and isolated patches of gravel, all over 200 feet O.D., which, apart from some doubtfully related patches near Worcester, fell into two groups: those in the main Severn valley between Gloucester and Tewkesbury, and those flanking the lower Leadon valley. The gravels of the first group consisted mainly of Bunter pebbles and were thought to be the outwash of a Severn valley glacier. Those of the second group contained both Llandovery and Malvernian rocks, and Wills therefore suggested that they had been brought down by a tributary issuing from the valley now occupied by the upper part of the Glynch Brook, which traverses the Silurian hills east of Ledbury and has one head-stream rising on Pre-Cambrian rocks. In Wills's opinion the two streams united near Gloucester (Wills, 1938, pp. 169–72, 219–20; see also Wills, 1950, fig. 28, in which, however, it is implied that the Leadon gravels are somewhat older than those of the Severn valley).

The present paper attempts both a re-examination of the gravels assigned by Wills to the Woolridge Terrace and an examination of certain high-level gravels west of the Severn valley which, though long known to exist, have never before been studied in detail. The paper is based upon a mapping of all gravel-patches over 200 feet O.D. within the quadrangle Ledbury–Tewkesbury–Gloucester–Huntley, together with analyses of their compositions. The results of the mapping are shown in Text-fig. 1.

II. Details of Gravels

The gravel-patches north and north-west of Upleadon have not been mapped before. They lie on a plateau composed mainly of Keuper Waterstones and Bunter Sandstones which, though deeply dissected by the Leadon and its tributaries, stands generally between 200 and 300 feet. The gravels occur on the divides between the streams, but it is

noticeable that large areas of the plateau stand somewhat higher than the base of the gravels and some areas even higher than their upper surfaces. This suggests that the present patches are the relics, not of a broad sheet of gravel, but of the flood-plain deposits of a shallow valley, perhaps no more than a mile wide, running from north to south across the present course of the Leadon.

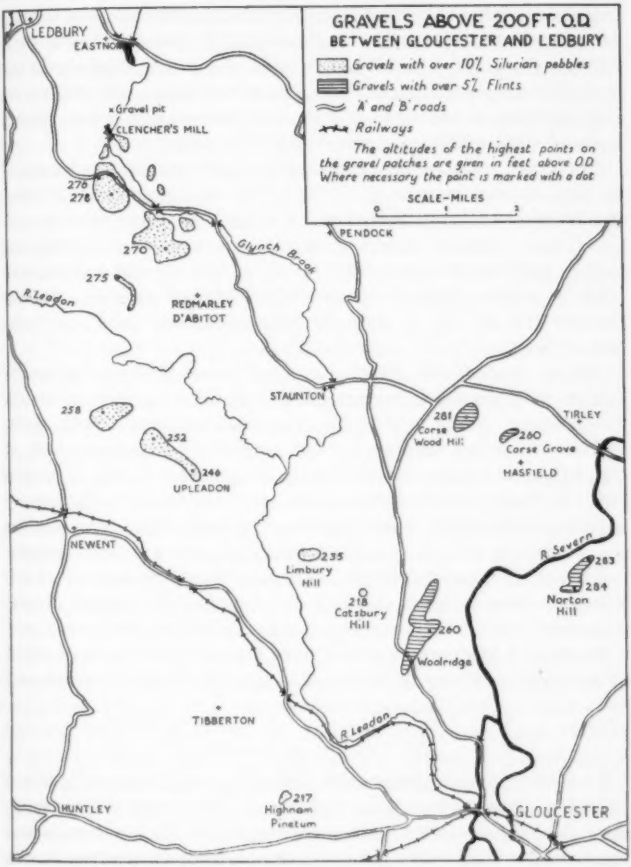
The south-eastern part of the area lies in the Severn valley proper, on outcrops of Keuper Marl, Rhaetic and Lower Lias. The ground here is mostly low-lying, but is diversified by a number of isolated steep-sided hills. Many of the hills rise above 200 feet, and it is on some of these, though not all, that the gravels occur. This is the area already mapped by Wills (1938, Pl. XV); the only gravel-patches not noted by him are those on Catsbury and Corse Wood Hills.

Nowadays these deposits are seldom well exposed. The only good section seen by the writer was a temporary one opened during road works in 1957 on the south-west side of A.417, 2½ miles south-east of Ledbury (733341). It showed a bed of red sand, 2 to 5 feet thick, with occasional Bunter pebbles and several lenses of pulverized coal ; there was marked current-bedding with all the laminae dipping to the south-east. Both above and below were extremely ill-sorted gravels containing Bunter pebbles and poorly rounded pebbles of Silurian rocks in about equal numbers, together with much sand and clay ; bedding was indistinct and highly irregular, indicating deposition in a torrential stream.

All the gravel-patches have been heavily eroded, and all, with the possible exceptions of those on Norton and Limbury Hills and the large patch ¾ mile south of Clencher's Mill, have lost their original surfaces ; this, of course, much reduces the significance of the levels shown on Text-fig. 1. The base of the gravels, also, is almost everywhere hidden by the gravels having themselves been moved downhill by solifluxion. Only two convincing lower contacts were seen, in road-cuttings near Bromsberrow Heath (733332, 2 miles north-west of Redmarley) and Compton Green (733288, 2 miles north-west of Upleadon), and these, though 3 miles apart, gave exactly the same level : 240 feet. It was also roughly estimated that the base of the gravel was at about 240 feet on Corse Wood Hill, 220 feet near Upleadon, and 195 feet at Highnam Pinetum. For Woolridge and Norton Hill, Wills (1938, p. 170) gave 240 and 250 feet respectively.

The compositions of the gravels at various localities had already been studied by Lucy (1872), Gray (1914), Wills (1938) and others, but it was decided to extend the examination to all the gravel-patches and to give it a quantitative basis. The procedure used was to clear an area of ground chosen at random, to sort the pebbles collected and then to calculate the percentage of each kind. Inevitably, of course, this

provided information only about the upper parts of each patch. As was expected all the gravel-patches showed a high proportion of Bunter pebbles; except in the extreme north-west this was usually between



TEXT-FIG. 1.—Map showing distribution of gravels above 200 ft. O.D between Gloucester and Ledbury.

80 and 90 per cent. Among the remaining pebbles, however, marked differences appeared. All the north-western patches had at least 10 per cent of Silurian pebbles, usually poorly rounded, together with

a few pebbles of Malvernian and basic igneous rocks (the latter probably from intrusions in the Cambrian near Eastnor) and very occasional flints and rolled Jurassic *Gryphaea*. Of the south-eastern patches Limbury, Catsbury, and Highnam Pinetum showed the same composition (the presence of Silurian pebbles on Catsbury was wrongly denied by Lucy, 1872, p. 81). At Woolridge, Norton Hill, Corse Wood Hill and Corse Grove, on the other hand, the gravels were almost free from Silurian pebbles but contained from 5 to 10 per cent of flints, the majority more or less brown-stained and either angular or only slightly rounded. These differences are shown in Text-fig. 1.

Two other gravel-patches near 200 feet were examined just south of the area covered by Text-fig. 1. These are on Barrow Hill (205 feet) and Hockley Hill (195 feet), 8 and 4 miles respectively south-west of Gloucester. Bunter pebbles were again predominant ; in addition Barrow Hill (mentioned in Wills, 1938, p. 171) showed 3 per cent of Silurian pebbles and two fragments of Malvernian rock, whereas Hockley Hill showed no definitely Silurian pebbles, but about 15 per cent of flints.

Among their minor identifiable constituents most of the gravel-patches have yielded a few pebbles of Cambrian quartzite, Carboniferous sandstone and pink or grey Carboniferous chert. Wills (1938, p. 170) found one pebble of Croft syenite, from Leicestershire, at Woolridge, and one pebble of volcanic ash, probably Welsh, at Norton Hill. At Woolridge, also, the present writer has found a single poorly rounded pebble of auto-brecciated rhyolite showing traces of spherulites and containing minute closely spaced quartz-veins; most probably this came from the Pre-Cambrian Warren House Series near Little Malvern. Gray's report (1914, p. 7) of pebbles of banded rhyolite (Uriconian type) at Highnam Pinetum has not been confirmed.

So far as is known no artefacts have been found in place in any of these deposits. The only records of indigenous fossils are mentioned on p. 165.

III. Origins of Gravels

It is clear from their compositions that the gravels shown in Text-fig. 1 were laid down by two separate streams. Their altitudes, however, show that the streams were not tributaries of one another, as was suggested by Wills. The present upper surfaces of the gravels with Silurian pebbles show a gradual decline from Clencher's Mill in the north to Highnam Pinetum in the south, and in spite of their advanced state of erosion it seems clear that the original surfaces must have done the same. It can hardly be doubted that these deposits form a single series ; the name proposed for them is the Upleadon Gravels. The four gravel-patches with abundant flints, on the other hand, are all

higher than the nearest patches of Upleadon Gravels ; Woolridge, for example, is 260 feet high and even the base of its gravels cannot be far below 240 feet, whereas Catsbury, only a mile to the west, does not rise above 218 feet. It is therefore concluded that these four deposits, besides having a different composition from that of the Upleadon Gravels, are also older ; it is proposed to give them a separate name, the Woolridge Gravels. It must be admitted that the two gravel-patches south-west of Gloucester do not fit easily into this scheme. That on Barrow Hill might perhaps be a patch of heavily diluted Upleadon Gravels, but that on Hockley Hill, with a level appropriate to the Upleadon Gravels and a composition appropriate to the Woolridge, remains unaccountable.

As for the ultimate sources of the deposits, it seems highly probable, though still not entirely certain, that the Woolridge Gravels are indeed the outwash of a glacier in the Severn valley. Their composition, however, suggests further possibilities which can best be considered after their dating has been discussed.

In the case of the Upleadon Gravels, both their composition and their distribution leave no doubt that Wills was largely correct in attributing them to a stream issuing from the upper Glynch Brook valley. There are, indeed, some indications of a tributary coming from the Severn valley and joining the main stream about a mile north-west of Redmarley: from here southwards the proportion of Bunter pebbles remains always between 80 and 90 per cent, whereas further to the north it is no more than 55 to 65 per cent ; moreover, numerous Bunter pebbles occur among the solifluxion deposits on the eastern slopes of the south end of the Malvern Hills, between 250 and 270 feet O.D. No signs of other tributaries have been found, however, whether from the upper Leadon valley or elsewhere.

The Glynch Brook valley itself, from Clencher's Mill to its head, is narrow and steep-sided, and little is visible on the surface except solifluxion deposits, yet it has provided one final piece of evidence in favour of Wills's suggestion. About 700 yards north of Clencher's Mill, on the west side of the valley, is the site, now completely overgrown, of the much-described gravel-pit at Clencher's Mill Wood (733356). The most recent account is by Gray (1914, p. 6), who describes 11 feet of sandy gravel with flints, Jurassic fossils, carbonaceous seams, and large sub-angular masses of local rocks ; this was overlain by 8 feet of sand with small pebbles, and the whole was covered by what seem to have been solifluxion deposits. The level given, 265 feet, probably refers to the base of the section ; the top of the sand would then be at 284 feet, which compares well with the level of 278 feet for the top of the Upleadon Gravels a mile to the south. The pit has yielded the worn molar of a mammoth (Symonds, 1883, p. 49) and a molar of *Elephas*

antiquus (Richardson, 1907, pp. 62–3); the former is stated to be in the Oxford Museum, but neither can now be traced.

Both Gray (1919, pp. 110–11) and Wills (1938, pp. 219–20) have suggested, somewhat tentatively, that the deposits here called the Upleadon Gravels may have been laid down by a stream running along the west side of the Malvern Hills and deriving its water partly from the Teme, partly from what is now the middle Severn valley. These possibilities are at present being investigated by the writer.

No useful guesses can be made concerning the sea-levels of these times. The average slope of the Upleadon Gravels is about 1 in 900, enough to suggest that the coast was still some distance away; the Woolridge Gravels cover so small an area that it is impossible to say whether they have any slope at all.

IV. Correlations

Both the Woolridge and the Upleadon Gravels are older than the Bushley Green Terrace, which in the Tewkesbury area does not rise above 175 feet O.D. (Wills, 1938, pp. 174–8). Since this is the equivalent of the oldest terrace in the Warwickshire Avon, No. 5, and since Avon Terraces 3 and 4 are generally assigned to the Last Interglacial and the beginning of the Last Glaciation (see Shotton, 1953, Table 6), the deposits must be at least as old as the Second-last (Gipping) Glaciation.

Beyond this it is impossible to go with certainty, but various speculations are suggested by the relatively high proportion of flints in the Woolridge Gravels. This was discussed by Wills (1938, p. 220), who reached no definite conclusion. The two most likely sources are the supposed outcrop of Chalk off the north Welsh coast, from which the flints of the Irish Sea Drift are thought to have been derived, and the Chalk outcrops of eastern England. Against the first alternative is the fact that as far north as Cheshire the Irish Sea Drift, according to information kindly supplied by Dr. I. M. Simpson, seldom contains as much as 5 per cent of flints, and there is no obvious reason why any earlier drift coming from the same direction should have contained more. There was, however, a period when abundant flints from eastern England were carried by ice into the south-west Midlands. This is represented by the Moreton Drift, a chalky boulder-clay lying in an embayment of the Jurassic escarpment only 25 miles east-north-east of Woolridge. Its stone-orientation indicates ice moving from north-north-east to south-south-west, which relates it to the Gipping advance (West and Donner, 1956, fig. 4). There is certainly no later chalky boulder-clay in the area; it is unlikely that there has ever been an earlier, for the Lowestoft Tills of central England appear to contain little or no flint (see Shotton, 1953, pp. 214, 240; West and Donner, 1956, p. 75) and their stone-orientations in any case denote ice moving

from the north-west, not from the east or north-east (West and Donner, 1956, fig. 4). It is therefore believed that the flints of the Woolridge Gravels were transported, at any rate for the greater part of their journey, by the glacier responsible for the Moreton Drift. This, incidentally, would account for the pebble of Leicestershire syenite found at Woolridge. It is no real objection that the Woolridge flints are usually brown-stained and patinated, whereas those of the Moreton Drift are fresh (Tomlinson, 1929, p. 162); flints would naturally be more subject to weathering and staining in a coarse gravel than in a clay.

It is unlikely that the flints completed their journey by being carried down by the Avon, as were those of the four younger terraces of the Severn, for at this time it is probable that the Avon, as a tributary of the Severn, did not yet exist (Shotton, 1953, pp. 249–50, 255), and in any case its mouth would almost certainly have been blocked by ice in the Severn valley. It is suggested, therefore, that the Moreton Drift glacier continued towards the south-west, over the sill of solid rock which is thought to have existed near Bredon Hill before the formation of the Avon (Shotton, 1953, pp. 248–9), and actually coalesced with the glacier in the Severn valley. The Woolridge Gravels, with both Bunter pebbles and flints, would thus be the outwash of the combined glaciers. Since the Moreton Drift belongs to the maximum of the Gipping glaciation and postdates the infilling of Lake Harrison (Shotton, 1953, pp. 242–3 and Table 6), so also, on this supposition, would the Woolridge Gravels, the Upleadon Gravels being later still.

A similar extension of the Moreton Drift glacier was suggested by Wills (1938, p. 222 ; 1950, p. 115), to account for thin deposits of stony clay found in a few places overlying the gravels of the Bushley Green Terrace; this, of course, implied that the Moreton Drift was much younger than the Woolridge Terrace. Wills himself, however, was doubtful of the glacial origin of these deposits (1938, pp. 175, 177, 179), which might perhaps be ascribed to solifluxion.

V. Conclusions

The oldest gravels of the lower Severn, probably of the entire Severn valley, are represented only by four small patches just north of Gloucester. It is proposed to call them the Woolridge Gravels. They consist mainly of Bunter pebbles, but also contain many flints. They could well be outwash gravel issuing from an ice-front perhaps a little south of Tewkesbury. It is suggested that the flints could be explained on the supposition that the Severn valley glacier was joined by a south-westerly extension of the glacier responsible for the Moreton Drift. This would make the Woolridge Gravels Gipping in age.

Slightly younger, but still somewhat older than both the Bushley

Green Terrace and the Last Interglacial, are certain gravels with a considerable extension to the north-west of Gloucester, here called the Upleadon Gravels. They also are chiefly composed of Bunter pebbles, but their special feature is the presence of much Silurian material. They were laid down by a stream issuing from the Silurian hills east of Ledbury, at a time when there appears to have been no important stream in the main Severn valley. The reasons for this state of affairs are still under investigation.

VI. Acknowledgments

The writer wishes to thank Professor F. W. Shotton, F.R.S., for much helpful advice, and Mr. and Mrs. A. N. Currer-Briggs, of Taynton, for their hospitality during work in the field.

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CORRESPONDENCE

ON SOME TURKISH ROCKS

SIR,—I have read with interest the letter by Ager (1) referring to what I may call boulder-beds, block-beds, and breccias of different types. The reader of that letter may well get the impression that our views on the origins of these rocks are confused like the rocks themselves. Let me refer to the Turkish examples first.

(*a*) Your readers will readily appreciate that many of the rocks mentioned in the first paragraph were described by Bailey and myself (2) in some detail. Although no reference was made to this work we are probably included in the group of " nappe-minded geologists ". However, the point here is that a *tectonic* origin was ascribed to the Ankara mélange.

(*b*) Although Turkish geologists have considered the Şile occurrences of mixed Cretaceous and Tertiary rocks as of tectonic origin also, I have suggested on a number of occasions in conversation with them that a *sedimentary* origin would better explain the phenomena. A summary of my views has been published (3).

(*c*) Reference was made to the joint paper by Tokay and myself on the Cretaceous deposits of the Zonguldak region. There we emphasized the *sedimentary-slump* nature of the rocks and pointed to the frequency of graded bedding and other features of turbidity currents. It is true that, at the time I read the paper, Migliorini compared the phenomena to those of the *Argille scagliose*. Since then other Italian geologists have, in conversation, compared the Ankara Mélange to the *Argille scagliose*. Accordingly the *Argille scagliose* is compared by the Italian School to both a tectonic mélange and to a sedimentary mixture. Recently Bailey and I have examined many sections of *Argille scagliose* in Tuscany and Liguria and we continue to be " nappe-minded ".

W. J. MCCALLIEN.

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March, 1958.

ANTIDUNES AND FLAME STRUCTURES

SIR,—It was with great interest that I read Lamont's further exposition of his views on the origin of " flame-structures", but I cannot agree with his explanation. It seems that Lamont envisages the development of an eddy of sand and water of sufficient violence to be able to support an overhanging thin film of clay. I cannot believe that such a structure could have more than a momentary existence; it would surely collapse once the eddy ceased. Even if such structures could be preserved, it would be wrong to attribute them to the action of antidunes. Gilbert (1914, p. 32) in his original description of antidunes, emphasizes that in them sand grains flow parallel with the bed in the troughs and up and down in the crests; he observes no reversed movement

and Bucher (1919, p. 165) clearly states that no vortices are involved. Thus the concavo-convex form of the flame-structure is not typical of antidunes. All authors agree that antidunes are symmetric in cross-section, and are "like very flat sine curves" (Blench, T., 1957, p. 8). Recorded wave-lengths and amplitudes are as follows (converted to cms.)

Wave-lengths around 23 cms. (Cornish, V., 1899, p. 626)
" " 450–600 cms., amplitude approx. 80 cms. (Pierce, 1916, p. 42)
" " 60–90 cms., amplitude approx. 15 cms. (Gilbert 1914, p. 32)

The figures given by Bucher seem to be an inaccurate conversion to metric units of those given by Gilbert. Many of the structures described by Lamont far exceed in relative amplitude anything yet described for modern antidunes. It is possible that the movement of the sand in the antidunes, since it does not appear to be by saltation, might generate movement in the partially consolidated layers below. This movement is known to occur below the larger "progressive sand-waves" described by Bucher. This dragging might produce flame-structures in an alternation of sand and clay, and it seems to me the only way in which antidunes could produce a flame structure.

Lamont attributes his antidunes to the action of vortices on a layer of sediment. Vortices appear to be generated on the bed of a modern stream only at water velocities associated with the formation of current-ripples i.e. at slower speeds than those which form antidunes (Bucher, W. H., 1919). In fact, Lamont's term "slow antidunes" seems self-contradictory, if by "slow" he implies movement of the order of 8–10 cms. pr. sec., as his introductory remarks suggest. If vortices are responsible for flame-structures, then they are more likely to be "dune-forming" than "antidune-forming".

Whilst it may be true that some flame-structures originated by vortical movement during deposition, in most cases a post-depositional origin is indicated as is clearly shown by Kelling and Walton (1957, p. 481 et seq.) Nevertheless I feel that the latter authors do not attach sufficient importance to "flow" (i.e. horizontal movement of sediment) as opposed to "loading" (i.e. vertical movement). Whilst agreeing that pure loading may explain many features, it is difficult to believe that flame-structures with a consistent direction of overturn do not reflect some horizontal movement of sediment. Whether such features should be called load casts or flow casts is arguable, since most examples involve movement which is neither purely horizontal nor purely vertical. Shrock's term "flow-cast" has priority of date (which commends it to the palaeontologist) whilst Kuenen's "load-cast" is now more commonly used. Nevertheless, there remains a category of structures which may have been *initiated* by flow of the overlying sediment, such as the structures I described (Prentice, 1956) from North Devon, and the examples of Group II of Kelling and Walton (1957, p. 485). The evidence presented by the latter authors for the origin of these structures as load-cast ripple-marks is not as yet convincing, and I look forward with interest to the promised expansion of these views. If we are to adhere to a genetic nomenclature for these sedimentary structures, it is important that the processes of formation should be properly understood before the nomenclature is applied.

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J. E. PRENTICE.

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4th February, 1958.

LOAD-CASTS AND FLAME STRUCTURES

SIR,—At the risk of over-inflating an already dangerously distended literature on load-casts and related penecontemporaneous sedimentary structures, may I add a comment to the interesting account of Kelling and Walton in the November-December issue of the *Geological Magazine* for 1957. It is unusual to conceive that ripple-mark should develop in muddy sediments, and if it did its presence would be revealed by current bedding, at least for transverse and interference-ripples. I have examined hundreds of perfectly displayed load-casts, and although current bedding is sometimes to be seen in the overlying sandstones, I have never seen it in the argillaceous under-beds. The authors have, however, reached a conclusion regarding the origination of mudstone injections from ripple crests that in all other ways agrees with my own, originally published many years ago (*Roy. Soc. Vict.*, Vol. 53, 1941, pp. 167–91) and referred to (with illustrations) more recently in *Outlines of Structural Geology* (London, 1953, pp. 11–13) including diapirism, analogy with salt domes (based in 1941 on Nettleton's experiments), and the suggestion that asymmetry of " flame " structures may reflect slight differential movements during compaction. In the Victorian examples, however, the ripple mark is actually preserved on a sandstone underlying the injected mudstone, not on the mudstone itself. The notion deriving from the later experiments of Parker and Macdowell that the height of the mud-domes may be controlled by the thickness of superincumbent sand has not escaped the writer, who was perhaps in error on this point in 1941, and it is interesting to note that the " overhang " so characteristic of salt domes is common in mudstone injections, which do in fact approximate to a definite elevation in any one bed (*Outlines*, fig. 6D, p. 11). The possible influence of differential loading (even after deposition amounting only to something of the order of an inch of sand) must not, however, be overlooked. True pseudonodules of the type erected by Marcar and Antun (*Bull. Soc. géol. Belg.*, vol. 73, 1950, pp. 121–49) most probably represent isolated or nearly isolated sand ripples with current-bedding (*Outlines*, fig. 6E) which sank into the soft mud developing flanges analogous with those of australites, due to fluid-dynamic moulding as they sank. Finally, I personally cannot resist the force of Sorby's beautiful illustrations and logic in ascribing flame structures and other mudstone injections in the Langdale slates, *without* any underlying ripples, to the breaking up of a deposit that was in a " creamy semi-liquid condition " (*Quart. Journ. Geol. Soc.*, vol. 64, pp. 171–233), a conclusion that agrees well enough with the concept, unknown to him, of turbidity current deposition and with experiments with oil films on water, which faithfully reproduce flame structures by differential flow. According to this notion asymmetrical flame structures are Helmholtz waves modified by the rheid properties of the materials so that the " anticlines " become pointed and inject the sandstone. The occurrence of examples with an amplitude of 6 to 10 feet in basalt resting on carbonaceous clay at Berwick, Victoria, is particularly interesting from this point of view. With sediments believed to be deposited by turbidity

currents it is more than a little difficult to decide what is to be regarded as "depositional" and what "post-depositional" in the way of small-scale structures, even with the one graded bed if this is complex, but, in the Victorian examples of salt-dome type mudstone injections, the beds (Silurian) are not strongly graded, but rather sharply defined alternating sandstone and mudstone. It was for this reason that the flame structures in them were regarded as post-depositional. Many examples in the Victorian Ordovician rocks, which are predominantly graded beds, are however to be ascribed to penecontemporaneous differential flow, as they are not related in any way to underlying ripples and there is widespread evidence for post-depositional flow of a few inches in the sandstones.

E. SHERBON HILLS.

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24th February, 1958.

SIR,—We regret having overlooked Professor Hills' important contribution to the subject of load-cast structures, but are glad that we have, however tardily, reached conclusions in certain respects similar to his own.

It is true that flame structures are perhaps best developed in purely argillaceous sediments, but it is the case, at least in the Southern Uplands, that the deformed bed may be of silt or even fine-grained greywacke. The formation of transverse ripple marks is thus not excluded though as we point out in our paper (p. 487) load-casts developed from them appear to be rare. In the example given by Professor Hills of flames developing in mudstone above transverse ripples, the original irregularities in the clay would reflect the transverse ripples and the structures may still be termed T-ripple load-casts.

The main aim of our paper, however, was to point out that given sufficient density contrast, any surface irregularity may be exaggerated to produce a load-cast: transverse ripple-mark is one of these irregularities, but grooves and flutes appear to be more important.

Differential flowage may produce flow-casts which are similar to load-casts, but so long as some of the original structure is retained the two forms can be recognized. In the rocks of the Southern Uplands we believe load-casting to be the dominant process but a general evaluation of their relative importance is an interesting task for the future.

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E. K. WALTON.

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10th March, 1958.

IGNEOUS ROCKS AT GOREI, SHILEMADU RANGE, BRITISH SOMALILAND

SIR,—May I add a postscript to Dr. J. E. Mason's interesting account (*Geol. Mag.* xciv, p. 498)?

Leckie's preliminary report on the geology of the Protectorate published in 1905 (1) seems to have been written whilst he was still in that country but presumably before he had seen the Gorei (8° 58′ N., 47° 48′ E.) exposure of which he makes no mention.

I have found no later published account by him, and the brief description of the geology of the country given in the Military Report on Somaliland, 1907 (2), does not refer to Gorei. But of the three attached maps, the two topographical sheets mark both Gorei and the Shilemadu Range, and the

geological map, " supplied by Mr. R. G. Edwards Leckie " and dated July, 1906, shows the whole Shilemadu Range as an isolated exposure of Archaean.

There appears to be no further record before I visited Gorei for an hour on 22nd March, 1930, and again, for a little longer, with Mr. J. A. Hunt, on 10th April, 1930. My field notes show that I felt undecided as to whether or not the igneous rocks were intruded into the Nubian Sandstone (Cretaceous), though the overlying Auradu Limestone (Lower Eocene, or possibly Palaeocene at base) was unaffected. My sketch shows conglomerate overlying the lower part of the igneous outcrop, and suggests that the Nubian overlay weathered igneous.

In the general description of the geology of the country (1933, p. 35) (3) the rocks were included in the Basement Complex assigned to the Archaean, since neither Mr. Hunt nor I had found evidence of any comparable igneous rocks of later age in the region. Specimens had by then been described, of both the crystalline rocks and the porphyry dykes, by Dr. Alfred Harker (1933, Appendix I, p. 43) and not by myself as implied by Dr. Mason.

My reference in 1952 (4) (p. 59) was made without a further visit and was influenced, so far as the dykes were concerned, by supposed analogy with Blanford's Cretaceous Trap Series, examples of which I had by then seen.

The nearest exposure of undoubted Basement Complex is found at Habarjeh, 100 miles N.W. of Gorei. The nearest Trap Series basalt is exposed in hand-dug wells at Warder, about 200 miles S.W.; and quartz tuffs, apparently of the same series, interbedded with the uppermost Nubian Sandstone, extend from near Warder to the N. and W., so far as Gudubi (8° 49′ N., 45° 00′ E.) in the Protectorate, and to the Harar Digit and Awareh (8° 16′ N., 44° 09′ E.) districts in Ethiopian territory (1952, Ill. 2, whose scale is 1 : 500,000 and not as stated).

As the result of Dr. Mason's paper I am content to accept his tentative view that all the igneous rocks of the Gorei exposure form part of the Basement Complex, a view that I should have thought his work makes firmer than he claims. Mr. Hunt tells me that, while he never managed t revisit Gorei, he also feels able to agree with Dr. Mason's findings.

Mr. Hunt's and my specimens are, it is believed, preserved in the Department of Mineralogy and Petrology at Cambridge.

W. A. MACFADYEN.

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HOPE'S GROVE,
TENTERDEN, KENT.

26th February, 1958.

REVIEW

MESURES GRAVIMÉTRIQUES ET MAGNÉTIQUES DANS LA PARTIE CENTRALE DE L'A.O.F. INTERPRÉTATIONS GÉOLOGIQUES. By YVONNE CRENN. Service de Documentation, O.R.S.T.O.M., 47 Boulevard des Invalides, Paris VII^{e}, 1957. The work comprises a brochure of 39 pages, size 10¾ in. by 8¼ in., with a folding map at the scale of 1 : 5 million, plus 3 separate maps at the scale of 1 : 1 million and 9 plates showing profiles. Price 3,000 francs.

During the past eight years important gravity measurements have been made in many parts of the world by French observers from expeditions organized by the Office de la Recherche scientifique et technique outre-mer (O.R.S.T.O.M.), a branch of the Ministère de la France d'Outre-Mer, and the Expeditions polaires françaises (missions Paul-Emil Victor). Because of the great care taken over calibration of instruments and to control the long-distance gravity connections between remote parts of the globe, these measurements, which have ranged from Greenland through Europe, Africa, Madagascar, and Australia to Terre Adelie, have made an important contribution towards the establishment of a consistent network of gravity base stations tied to a universal gravity datum. Interest in the gravity maps now published is heightened by the knowledge that they have been attached to this general framework.

The publication sets out the results of gravity and magnetic measurements made from 1953 to 1955 in an area of French West Africa comprising Côte d'Ivoire, Haute Volta, Togo, Dahomey, and parts of Soudan and Niger and goes on to discuss their geological interpretation. A first chapter describes the methods of measurement, reduction, and correction employed and discusses the precision obtained in the final Bouguer Anomaly values and Isostatic corrections. Though some of the Bouguer Anomalies depend on heights obtained by barometric levelling, the general accuracy appears to be about 3 milligals. The second chapter discusses general relations between the regional geology and the gravimetric and magnetic anomalies and the third and longest chapter deals with the fuller interpretation of the anomalies for individual districts and along specific profiles.

It is claimed that the relations between the geological structure and the gravity anomalies are examined by an original method which consists of comparing the gravity profiles obtained across all contacts of the same nature. Although this method is not particularly new, there is no doubt about its effectiveness in this application. In addition, the clarity of the discussion is greatly aided by the sensible scale of the profiles, presented as nine plates in the accompanying folder, and by the liberal provision of seventeen text-figures. The text itself is very condensed and in parts almost note-like in form, but these prove commendable virtues for, though the area to be described is large, yet a comprehensive survey of individual anomalies is achieved.

Utilizing the simplicity of the metric system, the gravimetric and magnetic profiles on the plates are drawn directly on graph paper with a horizontal scale of 1 cm. to 10 km., corresponding accurately to the 1 : 1 million scale of the three main Bouguer Anomaly maps. Each profile also carries an indication of the positions of geological contacts and values for isostatic corrections along the profile based on the Airy hypothesis for crustal thicknesses of 30 and 60 km. The use of graph paper and the scale of these profiles permit an unusually detailed examination of the results and, perhaps only because such an examination is possible, some minor criticisms or deficiencies may be noted. It would have been useful, for positioning the profiles against their corresponding maps, to have shown locations of the points of inflexion in the resolved lines of section. Profile A, Plate 1, provides a suitable example, for a series of gravity stations near the centre of the profile lie in a direction almost perpendicular to the overall direction of the section. Without a clear indication of how this change of direction has been projected, it is necessary to

consult the map carefully before it is possible to assess how far an apparent step in the gravity profile represents a transverse gravity gradient. On the same profile an apparent discrepancy between an anomaly of − 34 mgal. on the profile and − 26 mgal. on the map about 40 km. south of Parakou may be due to this slight positioning difficulty. Some indication of the topography along the profiles would have been of interest. The Bouguer anomalies have been derived assuming a uniform density of 2·67 to sea-level and some correlation of anomaly with topography may arise where the surface rocks have a diffierent density. Such effects will clearly be quite minor, however, in view of the density determinations listed and since large anomalies were encountered. Further, a geological map would have formed a valuable addition to the plates or alternatively a more detailed indication of the geological structure might have been shown on the Profiles. The main defect of the system adopted is that it fails to present any available geological information about the inclination of each contact but merely indicates the surface position. These suggestions, however, may well be regarded as counsels of perfection which themselves indicate the close interest aroused by the material presented.

As the author rightly points out, some of the anomalies encountered show a sufficiently promising agreement with the known geology to encourage further survey in greater detail and by additional methods. Two notable anomalies, a north-east to south-west gravity trough in Togo and an east to west gravity high which passes 20 km. north of Lomé, on the Dahomey coast, were followed to the eastern border of Ghana and knowledge of their further extent westwards is desirable. The publication of this material is therefore timely and lends interest to the gravity survey of Ghana recently commenced as part of Ghana's I.G.Y. programme. Finally, it is certainly to be hoped that future surveys of Ghana and Nigeria will be connected to the base-stations of the surveys now published, so as to provide eventually a continuous gravity picture along the lands north of the Gulf of Guinea.

W. B.

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